VOLCANO GEODESY: THE SEARCH FOR MAGMA RESERVOIRS AND THE FORMATION OF ERUPTIVE VENTS

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Abstract. Routine geodetic measurements are made at only a few dozen of the world's 600 or so active volcanoes, even though these measurements have proven to be a reliable precursor of eruptions. The pattern and rate of surface displacement reveal the depth and rate of pressure increase within shallow magma reservoirs. This process has been demonstrated clearly at Kilauea and Mauna Loa, Hawaii; Long Valley caldera, California; Campi Flegrei caldera, Italy; Rabaul caldera, Papua New Guinea; and Aira caldera and nearby Sakurajima, Japan. Slower and lesser amounts of surface displacement at Yellowstone caldera, Wyoming, are attributed to changes in a hydrothermal system that overlies a crustal magma body. The vertical and horizontal dimensions of eruptive fissures, as well as the amount of widening, have been determined at Kilauea, Hawaii; Etna, Italy; Tolbachik, Kamchatka; Krafla, Iceland; and Asal-Ghoubbet, Djibouti, the last a segment of the East Africa Rift Zone. Continuously recording instruments, such as tiltmeters, extensometers, and dilatometers, have recorded horizontal and upward growth of eruptive fissures, which grew at rates of hundreds of meters per hour, at Kilauea; Izu-Oshima, Japan; Teishi Knoll seamount, Japan; and Piton de la Fournaise, Réunion Island. In addition, such instruments have recorded the hour or less of slight ground movement that preceded small explosive eruptions at Sakurajima and presumed sudden gas emissions at Galeras, Colombia. The use of satellite geodesy, in particular the Global Positioning System, offers the possibility of revealing changes in surface strain both local to a volcano and over a broad region that includes the volcano.

INTRODUCTION

Much has been learned about the physics of active volcanoes by recording the slight surface displacements that occur before and during eruptions. Away from the eruptive site the surface rises or falls with geometric regularity as magma shoves its way through the crust. The pattern and rate of surface displacement can often be matched by a model that consists of a few simple pressure sources embedded in an elastic material. The displacement pattern reveals such characteristics as depth and rate of magma accumulation. Near the eruptive site, the crust exceeds its elastic limit, and the pattern of surface displacement is often chaotic as new faults and fissures form. Here it is difficult to interpret the action at depth because the displacement pattern depends partly on the sequence of ground breakage. Nevertheless, the net change in shape of the ground surface after an eruption reveals how magma was accommodated within the volcano. By multiplying the effects of a single event a hundred or a thousand times, we begin to understand how volcanoes grow.

In addition to rapid changes associated with individual eruptions, long-term changes occur in the shape of volcanoes. Hydrothermal systems that lie over many shallow magma bodies can cause the ground surface to rise or drop slowly. In addition, slow subsidence can be caused by the downward pull of gravity, causing a large volcanic mountain to spread. The net result of internal hydrothermal alteration or the spreading of a high volcano may be sudden collapse of the volcano, unrelated to magmatic activity, which is the major hazard at some long dormant volcanoes. The buildup to such events may be recorded by long-term measurement of surface displacement.

Historically, patterns of surface displacements were determined mainly by using high-precision land-surveying techniques, originally developed to make topographic surveys. These techniques measure changes in elevation, distance, or angle among a set of reference points called bench marks. Recently, techniques of space geodesy, such as the Global Positioning System (GPS), have been used to measure surface displacement at active volcanoes. Regardless of the technique, the strategy in making field measurements is simple. The relative positions are measured of a network of bench marks scattered across the surface. Some time later, the same points are remeasured to determine changes in relative position since the previous survey. A balance is sought between infrequent measurements of a large number of bench marks and frequent measurements of a few benchmarks. The first approach gives a detailed view of

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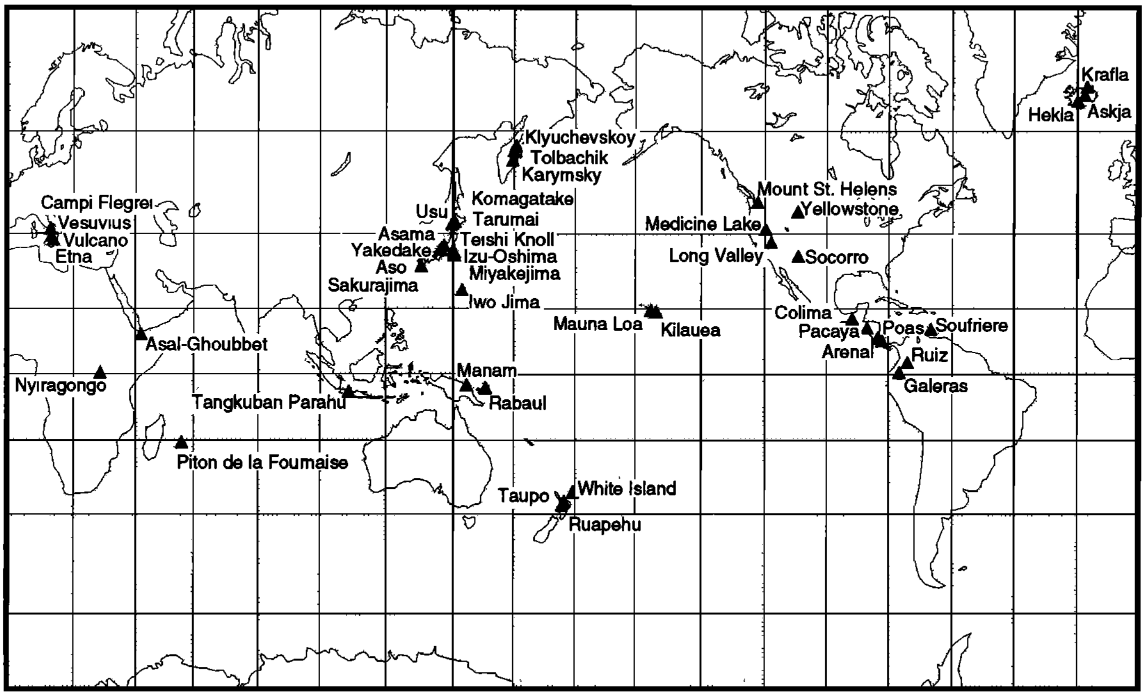


Figure 1. Locations of volcanoes are where measureable surface displacement has been reported. Geodetic measurements have been made at more volcanoes than shown here.

the shape of surface displacement, and the second reveals the timing of displacement.

Surface displacement has been measured at several dozen volcanoes worldwide (Figure 1 and Table 1), though only a half dozen or so have been studied intensely for more than a few decades. Often, measurements do not begin at a volcano until a crisis has developed, usually signaled by a shallow earthquake swarm, and so, in most cases, an unknown fraction of the total surface displacement associated with a volcanic event is recorded. It is at the half dozen or so volcanoes where measurements of surface displacement are routine that the most has been learned. Our discussion will concentrate on those few volcanoes. Before we examine how far or how fast the surface of an active volcano moves or describe the types of models proposed for the source of surface displacement, we need to examine extinct volcanoes whose insides have been laid bare by erosion or tectonics. We need to perform an autopsy on dead volcanoes to understand what gives life to the active ones.

VOLCANOES AS DYNAMIC LANDSCAPES

In the southernmost Andes, Pleistocene glaciers once cut deeply into the crust, scraping away surface features and exposing in geologic form a record of the events that occurred long ago within a now static crust. At several locations just east of the north-south line defined by currently active volcanoes along the Andes, the exposed rocks tell of a former comingling of mafic and silicic magmas, a relationship common in many sections of continental crust. The relationship is particularly clear at Cordillera del Paine, where glaciers laid bare a 2-km vertical section of a pluton that may have once fed a volcano [*Michael*, 1991].

The lower portion of the exposed pluton at Cordillera del Paine is a complex of gabbroic rocks. The remainder of the exposure is granite capped in a few places by older sedimentary rocks, making up the roof of the pluton. The composition of granitic dikes that invaded the roof suggests that the pressure at the roof of the pluton was about 200 MPa, which indicates an emplacement depth of 6–8 km [*Michael*, 1984]. The Cordillera del Paine pluton once occupied a magma chamber in the upper crust. Perhaps the pluton was the root of a volcano, though if such a volcano ever existed, all trace of it has been removed by erosion. At Cordillera del Paine we can only guess how the pluton may be related to volcanic activity. The importance of this site to us is that the silicic bodies that make up the pluton were invaded from below by basalt. Such a relationship supports the claim that basalt is the ultimate heat source powering all magmatic, and hence all volcanic, systems [*Hildreth*, 1981].

Magma Mixing

Other examples of silicic bodies invaded from below by basalt exist at the eastern extreme of North America. At Neward Island, Labrador, a layered intrusion records the mixing of mafic and silicic magmas [*Wiebe*, 1988]. At

TABLE 1. Geodetic Measurements of Active Magmatic Systems

Volcano	*Year Measurements Began*	*Most Frequent Measurement*	*Other Types of Measurements*	*References*
		Japan		
Usu, Hokkaido	1910	leveling	trilateration	*Minakami et al.* [1951], *Yokoyama et al.* [1981]
Tarumai, Hokkaido	1971	trilateration	spirit level tilt	*Yamashita et al.* [1984], *Mori and Suzuki* [1988]
Komagatake, Hokkaido	1929	leveling		*Mogi* [1958]
Sakurajima, Honshu (Aira caldera)	1891	leveling (annual)	triangulation, trilateration, tide gage, tiltmeter, extensometer	*Yokoyama* [1986], *Tada and Hashimoto* [1989]
Yakedake, Honshu	1977	spirit level tilt (annual)	trilateration	*Kimata et al.* [1982, 1989]
Asama, Honshu	1934	tiltmeter (daily)	leveling	*Minakami* [1942]
Teishi Knoll, Honshu	1968	leveling (annual)	trilateration, tiltmeter, dilatometer, GPS	*Fujita and Tada* [1983], *Okada and Yamamoto* [1991]
Aso, Kyushu	1937	leveling (decade)	trilateration	*Kubotera* [1984]
Miyakejima, Izu Islands	1940	leveling (annual)	tide gage	*Tada and Nakamura* [1988]
Izu-Oshima, Izu Islands	1979	leveling (~annual)	trilateration	*Hashimoto and Tada* [1990]
Iwo-Jima, Izu Islands	1976	leveling (semiannual)	trilateration	*Kumagai et al.* [1989]
		Russia		
Tolbachik, Kamchatka	1975		triangulation, trilateration, leveling	*Fedotov et al.* [1980]
Klyuchevskoy, Kamchatka	1978	leveling (annual)		*Fedotov et al.* [1990]
Karymsky, Kamchatka	1972	leveling (annual)	trilateration	*Magus'kin et al.* [1982]
		United States		
Mount St. Helens	1980	trilateration (annual), crack measurements	tiltmeter network	*Iwatsubo and Swanson* [1992], *Iwatsubo et al.* [1992], *Dzurisin* [1992]
Yellowstone	1976	leveling (annual)	trilateration	*Dzurisin et al.* [1994]
Medicine Lake	1988	leveling (semiannual)		*Dzurisin et al.* [1991]
Long Valley	1975	leveling (semiannual), tiltmeter, dilatometer	trilateration	*Savage and Clark* [1982], *Savage et al.* [1987], *Langbein et al.* [1993]
Socorro	1911	leveling (~20 years)		*Larsen et al.* [1986]
Kilauea	1919	summit tilt (daily)	leveling, spirit level tilt, trilateration, tiltmeter network, GPS	*Kinoshita et al.* [1974]
Mauna Loa	1965	trilateration (annual)	spirit level tilt, leveling, GPS	*Lockwood et al.* [1987]
		Mexico		
Colima	1982	leveling (annual)	trilateration, spirit level tilt, tiltmeter, GPS	*Murray* [1993]
		Central America		
Arenal, Costa Rica	1976	spirit level tilt (monthly)		*Wadge* [1983]
Poas, Costa Rica	1987	trilateration	spirit level tilt, GPS	*Rymer and Brown* [1989]
Pacaya, Guatemala	1979	leveling (annual)		*Eggers* [1983]
		Colombia		
Ruiz	1985	spirit level tilt (monthly), tiltmeter network	trilateration	*Banks et al.* [1990]
Galeras	1989	tiltmeter		*Lockhart et al.* [1990]
		Lesser Antilles		
Soufrière, St. Vincent	1977	spirit level (annual)		*Fiske and Shepherd* [1990]
		Iceland		
Krafla	1974	electronic tiltmeter (continuous), summit tilt (daily since 1976)	leveling, spirit level tilt, trilateration	*Johnsen et al.* [1980], *Tryggvason* [1994a]
Askja	1966	leveling (annual)	trilateration, lake level	*Tryggvason* [1989], *Rymer and Tryggvason* [1993], *Camitz et al.* [1995]
Hekla	1968	leveling (semiannual)	spirit level tilt, GPS	*Tryggvason* [1994b], *Sigmundsson et al.* [1992]

TABLE 1. (continued)

Volcano	*Year Measurements Began*	*Most Frequent Measurement*	*Other Types of Measurements*	*References*
		Italy		
Campi Flegrei	1819	leveling (monthly)	sea level, trilateration	*Berrino et al.* [1984]
Vesuvius	1974	leveling (semiannual)	trilateration, spirit level tilt	*Bonasia and Pingue* [1981]
Vulcano	1976	leveling		*Ferri et al.* [1988]
Etna	1971	leveling (semiannual)	trilateration, spirit level tilt, GPS	*Murray and Guest* [1982], *Nunnari and Puglisi* [1994]
		Africa		
Nyiragongo, Zaire (Congo)	1977	trilateration (semiannual)	spirit level tilt	*Kasahara et al.* [1982]
Asal-Ghoubbet, Djibouti	1972	leveling (decade)	trilateration	*Ruegg et al.* [1979], *Tarantola et al.* [1980]
		Indonesia		
Tangkuban Parahu, Java	1981	spirit level tilt (~annual)	trilateration	*Dvorak et al.* [1990]
		New Zealand		
White Island	1967	leveling (biannual)		*Clark* [1982]
Taupo	1979	leveling (~monthly)	lake level	*Otway* [1986]
Ruapehu	1970	trilateration (annual)		*Otway* [1979]
		Papua New Guinea		
Manam	1957	tiltmeter (daily)		*Mori et al.* [1987]
Rabaul	1973	leveling (~annual), tide gage, trilateration	spirit level tilt	*McKee et al.* [1985], *Archbold et al.* [1988]
		Réunion		
Piton de la Fournaise	1980	tiltmeters (continuous), extensometer (continuous)	trilateration, spirit level tilt, leveling, strainmeter	*Lenat et al.* [1989a, b]

the Isle au Haut Igneous Complex, Maine, a sequence of 10 alternating units of gabbro and diorite may represent the successive invasions of an andesitic magma body by a compositionally uniform mafic magma [*Chapman and Rhodes*, 1992]. Because the solidus of basalt is higher than that of a more silicic magma, a pulse of basaltic magma would superheat a silicic magma and cause convection. Vigorous convection would induce vesiculation, increasing magma pressure, perhaps to such a degree that the surrounding rock begins to fracture. If the fracturing extends to the surface, an explosive silicic eruption may ensue [*Sparks et al.*, 1977]. Such a basaltic trigger may have initiated the recent explosive eruption of Mount Pinatubo, Philippines [*Pallister et al.*, 1992].

Because basaltic magma occurs in nearly all magmatic provinces and because many large nonbasaltic eruptions are derived from zoned reservoirs, it is proposed [*Hildreth*, 1981, p. 10,153] that "nearly all magmatic systems are 'fundamentally basaltic' in the sense that mantle-derived magmas supply heat and mass to crustal systems that evolve a variety of compositional ranges." In the case of a large, old system, such as the volcanic systems beneath Yellowstone or Long Valley, episodic invasion of mafic magma is an efficient way to maintain a liquid silicic magma body [*Lachenbruch and Sass*, 1978]. We suppose that such episodic events occur now and cause surface strain as the crust accommodates a basaltic intrusion. The geologic record, however, does not indicate the frequency or duration of such episodic events. For that we rely on displacement measurements of active volcanic systems, such as those discussed in a later section, that may have recorded such events.

Development of Large Intrusions

The relation between large intrusive bodies and volcanic activity has been debated throughout the history of geology. Specifically, how does the crust accommodate emplacement of a large intrusion, whether as a series of sheets or as a large ellipsoidal body [*Pitcher*, 1979; *Paterson and Fowler*, 1993]? How might the emplacement affect the surface, even if the magma never erupts?

L. von Buch was among the earliest proponents of a "craters of elevation" theory for the formation of high volcanic mountains. His ideas were based on a visit to the Canary Islands in 1815, after which he noted that the filling of large, ancient magma reservoirs exposed in the eroded walls of the Canary Islands could have raised the volcanic islands. Modern investigations of the Canary Islands amplify von Buch's basic conclusion. Such volcanoes are built by both intrusive and extrusive processes, in roughly equal proportions [*Staudigel and Schmincke*, 1984]. At the island of La Palma, intrusions appear mostly as sills, which raised the overlying material and formed steep fault scarps on the flanks of the volcano. Upheaval is evident also at the basaltic islands of Rhum [*Brown*, 1956], Réunion [*Upton and Wadsworth*, 1969], and Saint Helena [*Daly*, 1927; *Baker*, 1968]. At these volcanic islands the large intrusive bodies consist of layered rocks that formed by episodic invasion of basalt [*Emeleus*, 1987]. At Saint Helena and Réunion the intrusive body is about 2 km thick and 10–20 km in diameter and lies about 2 km below the present surface. Around the edge of the body, individual intrusions form a plexus of crosscutting dikes and sills.

A contemporary twist of the "craters of elevation" theory has been recognized in the evolution of silicic calderas. Such a caldera forms by collapse during eruption of a voluminous ash flow tuff from a shallow magma chamber, probably fed from a shallow pluton. Repeated eruption and collapse of a single silicic caldera or a cluster of such calderas is related to the sequential development of shallow plutons, which coalesce to form a batholith [*Steven and Lipman*, 1976]. The association of a batholith and a silicic caldera is inferred from large negative Bouguer anomalies, characterized by steep marginal gradients, that enclose Yellowstone [*Eaton et al.*, 1975] as well as many other silicic calderas in the western United States [*Lipman*, 1984].

Growth of a batholith as a coalescence of individual plutons produces crustal uplift over two different scale lengths. The more easily recognized uplift is confined to the caldera floor and occurs within a few hundred thousand years after caldera collapse. The caldera floor may rise several hundred meters, a process called resurgence [*Smith and Bailey*, 1968]. Resurgence is most common at calderas larger than 10 km in diameter that formed in cratonic crust [*Lipman*, 1984], such as Valles caldera, New Mexico. Resurgence is less likely where the crust may be too weak to accumulate large volumes of magma, such as the Taupo volcanic zone, New Zealand [*Wilson et al.*, 1984], and in Japan [*Aramaki*, 1984].

Growth of the batholith may also produce regional uplift, which may occur over an entire volcanic field and be larger than any individual caldera. Such uplift is evident at the San Juan volcanic field, Colorado, as (1) a regional topographic high, (2) a slight tilting of volcanic strata away from the central part of the field, and (3) development of extensional structures across the volcanic field [*Lipman*, 1984]. In northern New Mexico, regional uplift is evident at the Sangre de Cristo Mountains, which are broadly arched over the granitic bodies associated with the Questa caldera [*Lipman*, 1983]. Such regional uplift involved the underlying batholith, which may continue to rise and intrude into the volcanic deposits that were erupted earlier from the growing batholith. Such a relationship is evident at the Casto batholith, which has intruded into the Challis volcanic deposits, Idaho [*Cater et al.*, 1973]. Because many batholiths are roofed by thin covers of their own volcanic ejecta [*Hamilton and Myers*, 1967], which obscure structural relationships in older rocks, field examples of regional uplift are difficult to notice.

Regional uplift might occur also if a volcanic system is supplied by an upwelling mantle plume, which provides dynamic support for a raised lithosphere [*Richards et al.*, 1988]. Rising plumes have been inferred to exist beneath a few dozen areas of hot spot volcanism, including Hawaii, Iceland, and Yellowstone [*Crough*, 1983].

Not all silicic calderas are associated with regional uplift. For example, regional subsidence occurred before formation of Long Ridge caldera, Oregon [*Rytuba and McKee*, 1984], and occurs now around Campi Flegrei caldera, Italy [*Dvorak and Mastrolorenzo*, 1991], and Medicine Lake caldera, California [*Dzurisin et al.*, 1991]. In each case, subsidence may be a combination of regional extension, gravitational loading of the crust by erupted material and depressurization of a hydrothermal system.

Geophysical Evidence of Shallow Magma

In this and the next subsection, we digress from a consideration of extinct volcanoes and consider what geophysical evidence indicates the presence today of magma in the shallow crust. *Lipman* [1984] noted that the large negative Bouguer gravity anomalies associated with many extinct silicic calderas probably indicate shallow batholiths. Such a Bouguer anomaly is also associated with Yellowstone caldera, where the anomaly may define a shallow cooling granitic pluton that still contains partial melt [*Eaton et al.*, 1975]. The region beneath Yellowstone caldera has a lower seismic velocity than its surroundings and does not everywhere transmit shear waves; both seismic characteristics are possible evidence of a shallow body of partial melt. How much of the region beneath Yellowstone caldera does other geophysical evidence indicate is molten today?

Magma has supplied eruptions of the Yellowstone volcanic system for at least the last 2 m.y., most recently 75,000 years ago, and so magma is probably present now [*Hildreth et al.*, 1991]. A three-dimensional model of seismic velocity beneath Yellowstone shows a large low-velocity zone about 100 km in diameter that extends from the surface to a depth of about 250 km [*Iyer*, 1988]. The velocity reduction is about 10% throughout most of the low-velocity zone, and the reduction is as much as 30% in two small areas within 10 km of the surface beneath the northeast and southwest sections of the caldera [*Lehman et al.*, 1982]. Major changes in temperature and composition across the caldera boundary, without the presence of magma, could account for the 10% reduction of velocity [*Lehman et al.*, 1982]. The 30% velocity reduction, however, does require a partial melt at depths greater than about 2 km, with a melt fraction that ranges from 10 to 50% [*O'Connell and Budiansky*, 1977]. At depths shallower than 2 km, the velocity reduction is probably caused not by partial melt but instead by a large volume of steam in a highly porous material [*Lehman et al.*, 1982].

By analogy with Yellowstone, the 30% reduction of seismic velocity beneath Valles caldera, New Mexico, is also evidence of a body of partial melt [*Roberts et al.*, 1991]. The body is about 17 km across, is centered at a depth of 10 to 13 km, and must have a vertical height of at least 8 km. As at Yellowstone, in order for melt to still exist beneath Valles caldera and to account for the present heat flow, the Valles pluton must have been heated throughout most of its period of activity, probably by intrusion of basalt.

Elsewhere, geophysical evidence is not as compelling for the presence of shallow magma, even though continued eruption of some volcanoes indicates that shallow magma must be present. An aseismic, low-velocity zone beneath the summit area of the basaltic volcano Kilauea, Hawaii, represents a zone of temporary magma storage [*Thurber*, 1984]. Because the velocity reduction is only a few percent, the zone of magma storage is mostly hot solid rock that contains pockets of magma. Clear identification of magma storage is complicated by high-velocity zones within Kilauea, which represent dense intrusions [*Rowan and Clayton*, 1993]. On the basis of the pattern of movement of Kilauea's south flank since a magnitude 7.2 earthquake in 1975, *Delaney et al.* [1990] proposed that a magma storage zone extends from 3 to 9 km beneath the summit region and both of Kilauea's rift zones. The existence of deep magma storage beneath the rift zones based on geodetic data, however, is questioned because Delaney et al. failed to consider the response of Kilauea to the magnitude 7.2 earthquake and the contribution to surface strain caused by a decade of aftershocks [*Dvorak et al.*, 1994].

At Krafla, Iceland, another basaltic system, magma storage is inferred from attenuation of shear waves [*Einarsson*, 1978]. The attenuation zone is about 2 km wide, lies within 3 km of the surface, and extends no deeper than 7 km. A similar zone may represent magma storage less than 3 km beneath the basaltic volcano Plosky Tolbachik, Kamchatka [*Balesta et al.*, 1978].

Beneath Long Valley caldera, California, the velocity reduction is only about 10 per cent [*Steeples and Iyer*, 1976], and so is not an unambiguous indicator of a partial melt. Instead, attenuation of shear waves in two small zones [*Sanders*, 1984] and refraction of seismic waves [*Hill*, 1976] are better indicators of possible shallow magma storage.

Results of seismic reflection surveys are used to infer structures that could be the tops of small magma bodies. Such a body lies about 4 km beneath the floor of Campi Flegrei caldera, Italy, at the base of an intense zone of shallow earthquakes [*Ferrucci et al.*, 1992]. A 20-km-deep body lies beneath Etna [*Sharp et al.*, 1980]. Results of other reflection surveys reveal a sill-like body, about 8 km wide and less than 0.5 km thick, about 20 km beneath Socorro, New Mexico, along the Rio Grande Rift Zone [*Sanford et al.*, 1977; *Ake and Sanford*, 1988]. Continuous seismic profiles across several segments of the mid-ocean ridge system reveal the top of crustal magma chambers 1–2 km below the seafloor [*Mutter*, 1985; *Detrick et al.*, 1987; *Collier and Sinha*, 1992].

No evidence of either low-velocity or attenuation zones has been found beneath isolated andesitic or dacitic volcanoes in the Cascade Range, in particular Mount Hood, Oregon, and Mount Shasta and Mount Lassen, California [*Iyer*, 1988]. The lack of evidence means that magma rises from a deep source through narrow conduits that are smaller than can be detected by seismic methods. A small magma body is inferred from seismic studies to lie about 5 km beneath the summit caldera of the mixed basaltic and rhyolitic Medicine Lake volcano in California [*Evans and Zucca*, 1988]. Seismic studies are also used to infer the presence of a shallow layer of boiling water at Medicine Lake [*Evans and Zucca*, 1988] and Newberry caldera, Oregon [*Zucca and Evans*, 1992].

In summary, except for bodies beneath Yellowstone and Valles calderas, geophysical evidence is consistent, but not conclusive, for the presence of magma in the upper crust beneath basaltic volcanoes. Nonetheless, small, shallow magma bodies are inferred to exist beneath many volcanoes where eruptions are frequent and hydrothermal activity is conspicuous. We will show that the location of these bodies often corresponds to the location of buried pressure sources determined by measurements of surface displacement.

Hydrothermal Systems

Vigorous fumaroles, geysers, and boiling springs are often associated with an active volcanic system. At such a system the action of hydrothermal fluids may contribute to the pattern of surface displacement, either amplifying or suppressing movement caused solely by magmatic processes. The Yellowstone system is used as an example because scientific drilling and intensive geochemical and isotopic studies of hydrothermal fluids have characterized the detailed extent of that system.

At Yellowstone, meteoric water circulates under hydrostatic pressure to depths as deep as 5 km [*White et al.*, 1975; *Fournier*, 1989]. Below 5 km, hydrothermal circulation is choked off by reduced permeability because earthquakes are not frequent enough to reopen fractures that became clogged by mineral deposition. Thus a self-sealed layer develops that separates circulating water at hydrostatic pressure and deeper fluids at lithostatic pressure. The latter includes fluids released during cooling and crystallization of magma. The presence of a self-sealed layer near the hydrothermal-magma interface could confine temporarily a pressurized aquifer. The cyclic increase and release of pressure (the beginning of the latter signaled by an earthquake swarm) raises and then lowers the ground surface.

The existence of such a self-sealed layer is confirmed by results of intensive drilling at the active caldera of Campi Flegrei, Italy, and into a fossil hydrothermal system associated with the Henderson porphyry molybdenum deposit in north central Colorado. Drilling at Campi Flegrei encountered an impervious 1-km-thick layer that formed by hydrothermal alteration [*Rosi and Sbrana*, 1987]. Below the layer is a confined aquifer at temperatures that exceed 250°C.

In Colorado, drilling and underground geologic mapping revealed a system of radial, mineral-filled fractures that surround a molybdenum deposit, which owes its existence to the upward flow of fluids from a former magma body [*Carten et al.*, 1988]. The radial fractures formed when local pressure exceeded lithostatic pressure, perhaps caused by dilation as fluids exsolved from a crystallizing magma or as new magma intruded into a preexisting magma body. Such dilation almost surely raised the ground surface. The surface may lower if the self-sealed layer ruptures temporarily, fluids escape, and the pressure of the remaining fluids drops.

Another indication that strain changes caused by a magmatic system can influence the circulation of ground fluids is the change in water level in deep wells. The water level of a 370-m-deep well 2 km east of Usu volcano, Japan, rose almost 40 m after the beginning of the August 1977 eruption [*Watanabe*, 1983]. The water level continued to rise for another year, approximately the duration of the eruption, then dropped and returned to its preeruption level 3 years later. The sudden rise during the yearlong eruption, when the amount of surface displacement was 100 m or more, suggests that the water level responded to compression of the aquifer as magma rose toward the surface. The subsequent drop in water level followed an exponential decay with a decay constant of 3 years, the slower diffusion of water driven by increased pore pressure.

Water level responded in a similar way to strain pulses along the segment of the San Andreas fault system near Parkfield, California [*Roeloffs et al.*, 1989]. The strain pulses, recorded by creep meters, were sometimes accompanied by a sudden change in water level of a confined, 250-m-deep well. The water levels recovered in 1–2 months and, as at Usu volcano, can be modeled as diffusion of water by increased pore pressure. These examples illustrate a possible way to distinguish magmatic and hydrologic processes. A magmatic process can begin within a day or less. The hydrologic response to a strain change is more gradual, lasting months or even years.

Sills, Laccoliths, Plugs, and Dikes

So far, we have considered those processes that cause relatively slow surface displacements over periods of several months or longer. Sudden displacements, which may last less than an hour, are also recorded at many volcanoes, often immediately before an eruption. The geology of eroded volcanoes gives ample illustrations of features with discordant or chilled contacts that formed rapidly. A short encyclopedia of such geologic forms and structures is presented by *Frankel* [1967]. We review the dimensions of four general volcanic forms: sills, laccoliths, plugs, and dikes.

Sills are horizontal, sheetlike bodies, often injected along a well-defined geologic contact or between thinly bedded units. The thickness of a sill may be remarkably uniform across an exposure of 10 km or more. The range of thicknesses of different sills is from about a meter to a few hundred meters. A sill, like the other intrusive bodies described here, may form by successive injections of magma. Because the horizontal geometry of a sill does not intersect the surface, a sill is usually considered only as a means of storing magma, perhaps temporarily. The geometry of joints in a parallel set of sills at Katmai, Alaska, shows that sills are also a means of transporting magma horizontally [*Lowenstern et al.*, 1991]. Horizontal transport via a sill may have occurred during simultaneous eruptive activity at vents separated by 10 km during the 1912 eruption of Katmai that formed the Valley of Ten Thousand Smokes [*Hildreth*, 1987].

If horizontal extension of a sill is retarded, so that continued injection of magma causes vertical growth of the intrusive body and arching of overlying layers, then the sill evolves into a laccolith. Laccoliths are less common than sills, though they are not rare; more than 600 have been described in the United States [*Corry*, 1988]. Field observations suggest that sills are the forerunners of laccoliths, which form only after magma has spread far enough laterally to gain leverage to bend the overlying strata upward [*Johnson and Pollard*, 1973]. On the basis of cooling rate, the time required for a sill or laccolith to form within cold rock is less than a few years [*Corry*, 1988].

A volcanic plug, also called a "volcanic neck," is a vertical, pipelike body of magma that may be the conduit of a former volcanic vent. Such a feature has no exposed or inferred floor. Instead, a plug connects to a lower magma body that was the source of eruptions. A clear example of this field relation is exposed near the Cloudy Pass batholith, Washington [*Cater*, 1960]. The southeastern end of the batholith plunges beneath older crust that is pierced by a line of plugs.

A dike is a thin intrusive body with a vertical or nearly vertical orientation. Often, dikes are crowded in swarms, either as a radial pattern around a central point or along a long, linear system. Linear systems are considered separately in the next subsection. The central point of a radial pattern may be a complex of plugs or laccoliths, such as at Sunlight volcano, Wyoming [*Parsons*, 1939] and the island of Rhum [*Brown*, 1956]. Radial dikes on the dormant volcano Mount Rainier, Washington, radiate from the summit that overlies a central plug [*Fiske et al.*, 1963]. A radial dike pattern may result from the tensions produced when pressure increases within a central conduit, such as a plug. The expansion causes the surrounding rock to split in vertical cracks that extend outward from the conduit [*Iddings*, 1914], a process analogous to hydraulic fracturing in oil wells [*Hubbert and Willis*, 1957]. A regional stress field or heterogeneous crust may reorient a radial dike pattern into a broad, linear system far from the central point, such as Spanish Peaks [*Ode*, 1957].

Other dike geometries are also possible, principally ring dikes or cone sheets, which have arcuate or closed circular outcrops. Ring dikes are vertical or inclined outward slightly from a central point. Cone sheets are inclined inward. From the different orientation with respect to a central point, which is the buried position of a large plug, ring dikes are the result of subsidence and cone sheets are the result of expansion [*Anderson*, 1936]. The emplacement sequence of cone sheets and ring dikes is usually complex, such as on the island of Rhum, which indicates repeated episodes of expansion followed by collapse over a central magma body. The association of ring dikes with an ash flow eruption of the Questa caldera, New Mexico [*Lipman*, 1984], suggests that some ring dikes may develop during the catastrophic events that form silicic calderas.

Linear Dike Swarms

In Iceland, erosion has exposed thousands of dikes aligned parallel to the axis of the mid-ocean ridge system. In a survey of 700 basaltic dikes in eastern Iceland, *Gudmundsson* [1983] determined an average width of about 4 m and followed the longest exposed dike for more than 22 km. In Hawaii, basaltic dike widths measured at the eroded Koolau volcano range from 0.05 m to 6.7 m; the average width is 0.7 m [*Walker*, 1987]. At Hakone volcano, Japan, the widths of andesitic dikes are mostly between 1 and 5 m, and the average thickness is about 3 m [*Kuno*, 1964]. These comparisons suggest that dike width is determined by factors other than magmatic composition. This is supported by field studies at sites where contemporary dikes of different compositions formed. For example, in the Lesser Antilles, basaltic and andesitic dikes of similar age and tectonic setting have a similar spectrum of dike widths [*Wadge*, 1986]. If all other factors are equal, fracture mechanics suggests that the average width of a dike increases with dike height or length and that the ratio of average width to length is proportional to the ratio of magma driving pressure to host rock stiffness [*Pollard*, 1987].

A comparison of recent volcanic activity at Krafla, Iceland, and Kilauea, Hawaii, suggests that more dikes reach the surface in Hawaii than Iceland [*Rubin*, 1990]. The difference may be caused by the steady regional extension across Iceland, while a relatively high magma supply to Kilauea maintains compressive stress on the flanks of Kilauea. Dikes in Hawaii that vent usually do so along only a part of their total length [*Casadevall and Dzurisin*, 1987].

Flow lineations of dikes show that magma moved outward and upward from a central plug at Spanish Peaks, Colorado [*Smith*, 1987], and Otoge volcano, Japan [*Takada*, 1988]. Downward projection of lineations at Koolau volcano, Hawaii, indicate that magma rose in dikes from a center now 2–3 km below sea level, probably the top of a magma reservoir [*Walker*, 1987].

Exposed dikes extend as far as 35 km from central plugs at Spanish Peaks [*Smith*, 1987] and Otoge volcano [*Takada*, 1988], which suggests horizontal transport of magma of at least that distance at these central volcanic systems. During historical eruptions, simultaneous activity along rift systems suggests horizontal magma flow over distances of at least 50 km in Iceland [*Sigurdsson and Sparks*, 1978] and Hawaii [*Eaton and Murata*, 1960]. Horizontal flow of hundreds to thousands of kilometers occurred during emplacement of dikes into the lower crust from large basaltic centers, such as the Mackenzie swarm, Canada [*Ernst and Baragar*, 1992], and the Mull swarm, Scotland [*Macdonald et al.*, 1988]. Because an individual dike of these swarms was emplaced by a single pulse of magma, the horizontal growth rate of a dike must have been at least a few kilometers per hour [*Macdonald et al.*, 1988].

HOW FAR AND HOW FAST?

Compared with familiar geologic standards, the amount and rate of surface displacement can be astounding at an active volcano. A meter or more of displacement recorded in a day or less is not unusual. Smaller amounts and rates are also common, sometimes over an area of a hundred square kilometers or more. Such a wide range of possible surface displacement illustrates the difficulty in making measurements. To make useful measurements, techniques must have (1) a broad range to ensure that an entire event is recorded, (2) sufficient sensitivity to record slight and slow changes, and (3) adequate spatial coverage to capture an event. A single technique cannot satisfy all three requirements, and so multiple techniques are often used. Detailed discussions of survey errors and accuracies and of the installation and stability of benchmarks are given by *Ewert and Swanson* [1992] and *Murray et al.* [1994]. A description of the installation and operation of tiltmeters is given by *Toutain et al.* [1994]. In this section we illustrate successful techniques by showing specific examples of the amount and rate of displacement.

Long-Term Displacements

A history of displacement longer than a few decades is available for only a few volcanoes. We concentrate on the four examples that have the longest record of measured surface displacement; three are calderas that formed by large explosions several thousand years ago and the fourth is an intraplate basaltic volcano. Two of the four have erupted since measurements of surface displacements were begun. We augment this discussion with examples from several other volcanoes where measurements were made at least a few years before an eruption.

Sakurajima. Sakurajima is an andesitic-basaltic cone on the southern edge of Aira caldera, which forms Kagoshima Bay on Kyushu Island, Japan (Figure 2a). Of the many eruptions of Sakurajima during the 1200 years of historical records, one of the largest was the 1914

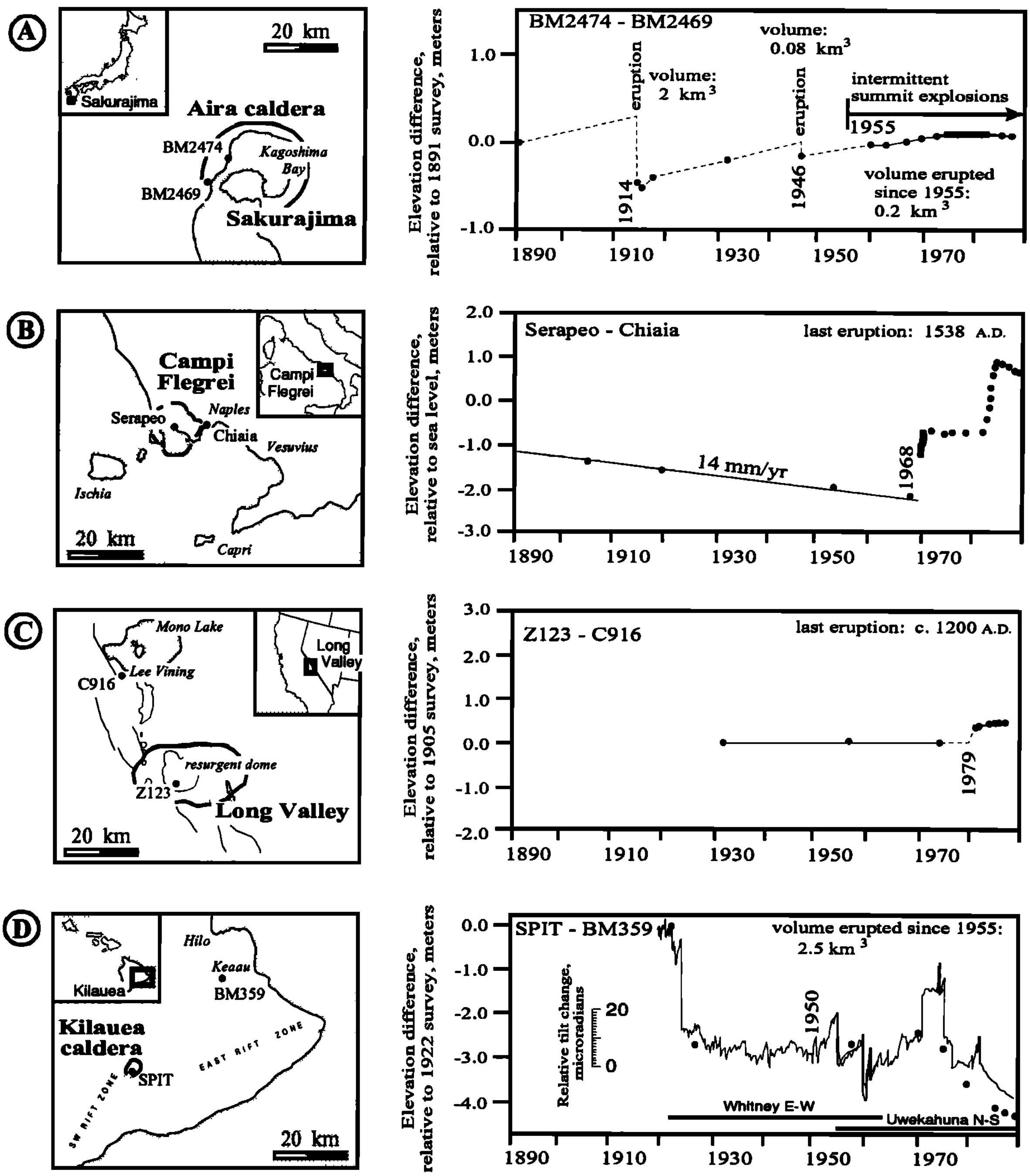


Figure 2. Long-term vertical movement at four active volcanic systems. Caldera outlines are shown on each map as thick solid lines. The time series plots to the right are relative elevation changes based on leveling surveys; bench mark locations are shown in the maps on the left. (a) Sakurajima. This andesitic-basaltic volcano, located on the southern rim of Aira caldera, has been the sole eruptive site of the caldera since at least the eighth century A.D. [*Yokoyama*, 1986]. The dashed line in the time series plot is an inferred history of vertical movement [*Sassa*, 1956]. The vertical scale of elevation change is twice that shown for the other three volcanoes because bench mark BM2474 is about 5 km from the presumed spot of maximum vertical movement in the caldera center (Figure 3a). Bench marks for the other calderas are within several hundred meters of the spot of maximum vertical movement. (b) Campi Flegrei. From at least 1820 until 1968, the caldera center subsided at an average rate of 14 mm yr^{-1}. Since 1968, uplift of as much as 3 m has occurred in two distinct periods, the first ending in 1972 and the second lasting from mid-1982 to December 1984 [*Berrino et al.*, 1984]. (c) Long Valley. Though the last pre-1980 leveling survey was conducted in 1975, trilateration surveys across the eastern section of the caldera suggest that no significant ground movement occurred until after 1979. Since 1980, uplift rate has been continuous, though variable (as an example, see Figure 19b) [*Langbein et al.*, 1993]. (d) Kilauea. Since 1919, bench mark SPIT subsided more than 4 m. A detailed history of vertical changes is inferred from daily tilt measurements, shown as a thin solid line.

eruption, when 2 km^3 of lava was erupted. Using the results of an 1891 leveling survey and a repeat survey in 1914 after the eruption, *Omori* [1918] showed that the edge of the caldera subsided at least half a meter. The subsidence pattern had the shape of a shallow bowl (Figure 3a). Maximum subsidence was not centered around the 7-km-long eruptive fissure that formed across Sakurajima, but, instead, was located at the center of Aira caldera. The difference between triangulation surveys conducted before and after the eruption showed as much as 8 m of crustal extension across the eruptive fissure [*Mogi*, 1958].

A 1915 leveling survey showed that subsidence continued at least a few months after the eruption [*Yokoyama*, 1986]. The caldera floor began to rise by at least 1918 and continued to rise until the next eruption in 1946 (Figure 2a). After a decade of dormancy, intermittent explosions began in 1955 from the summit crater of Sakurajima and continue today. About 0.2 km^3 of ash was erupted through 1982 [*Yokoyama*, 1986]. Frequent leveling surveys since 1961 show that the caldera floor rose slightly until about 1979, after which it subsided at a rate of a few millimeters per year [*Tada and Hashimoto*, 1989].

Campi Flegrei. In contrast to the frequent eruptions of Sakurajima, Campi Flegrei, on the west coast of southern Italy near Naples, has erupted only once in 4000 years, an explosive eruption that lasted 10 days in A.D. 1538. Frequent leveling surveys since 1905 reveal a history of vertical displacements at Campi Flegrei that is different from that at Sakurajima and Aira caldera. Between 1905 and 1968 the center of the Campi Flegrei caldera subsided at a rate of 14 mm yr^{-1} [*Dvorak and Berrino*, 1991] (Figure 2b), essentially the same as the average subsidence rate determined from sea level rise at Campi Flegrei since the 1820s [*Berrino et al.*, 1984]. Between the summer of 1968 and January 1970, the caldera center rose almost a meter and continued to rise at a decreasing rate until mid-1972 [*Berrino et al.*, 1984]. No earthquakes were widely felt during the uplift. The caldera subsided during the next 3 years, then showed no elevation changes until mid-1982. A second episode of uplift began in mid-1982 and continued at a steady rate until it ended suddenly in December 1984. The net uplift during the two episodes since 1968 was 3.0 m. The second episode was a period of intense earthquake swarms, which damaged many buildings and caused a temporary evacuation around the caldera center [*Barberi et al.*, 1984]. The center of the caldera floor has subsided at a rate of 10–30 mm yr^{-1} since December 1984.

An earlier record of vertical displacement is available at Campi Flegrei based on historical documents and on examination of Roman ruins. Three marble columns at a Roman marketplace built in the second century B.C. near the caldera center, which were above sea level when the marketplace was excavated in 1750, are perforated by marine shells. This is clear evidence that since the Roman Age the caldera center had subsided several meters, submerging the marketplace beneath the Mediterranean Sea, then was raised by a similar amount [*Dvorak and Mastrolorenzo*, 1991]. Government documents and eyewitness reports show that uplift of the caldera center began at least a few decades before the one historical eruption in A.D. 1538 and accelerated during the 2 days before the eruption.

Long Valley. As at Campi Flegrei, sudden uplift occurred recently at Long Valley caldera, California. Long Valley is considerably larger, 17 by 32 km versus 12 km, and older than Campi Flegrei. Long Valley caldera formed about 700,000 years ago, while Campi Flegrei formed about 32,000 years ago and had a second major collapse about 11,000 years ago. The most recent eruptions of Long Valley and Campi Flegrei occurred several hundred years ago. The A.D. 1538 eruption of Campi Flegrei was near the caldera center, whereas a series of eruptions near Long Valley occurred in about A.D. 1200 along the Mono volcanic chain, an alignment of dacitic domes and craters that intersects the western rim of Long Valley caldera.

The results of leveling surveys at Long Valley show that little if any vertical displacement occurred between 1932 and 1975 [*Savage and Clark*, 1982] (Figure 2c). Trilateration measurements across the south portion of the caldera suggest that no significant displacement had occurred by July 1979. In May 1980, four $M > 6$ earthquakes beneath the southern caldera boundary may have signaled the beginning of unrest. The next leveling survey in September 1980 showed that the caldera center had risen 0.25 m since 1975. Since 1980, results of frequent leveling and trilateration surveys indicate continued rise of the caldera center at variable rates, the highest rate during January 1983 [*Castle et al.*, 1984] and for several months beginning in October 1989 [*Langbein et al.*, 1993]. The total uplift since 1980 now exceeds 0.6 m.

Other explosive calderas, where measurements have not been as frequent or do not have as long a history as the three already described, also have sudden episodes of inflation, indicated by distension or uplift. For example, Usu volcano on the southern edge of Toya caldera, Hokkaido, Japan, showed sudden horizontal extension of 0.24 m less than 2 hours before the start of the 1977 eruption, though no comparable displacement rates were measured during the previous 22 years [*Watanabe*, 1983]. The floor of Rabaul caldera, Papua New Guinea, rose almost 2 m between 1973 and 1984; uplift was steady until 1982, then increased suddenly and was accompanied by intense swarms of shallow earthquakes [*Archbold et al.*, 1988]. The floor of Yellowstone caldera rose 0.7 m between 1923 and 1975, though the history of this rise is unknown because no intervening surveys were conducted [*Pelton and Smith*, 1982]. Annual surveys since 1983, however, showed 1 year of slow uplift at a rate of about 20 mm yr^{-1}, followed by 1 year of no elevation change, then several years of slow subsidence at an average rate of about 20 mm yr^{-1} [*Dzurisin et al.*,

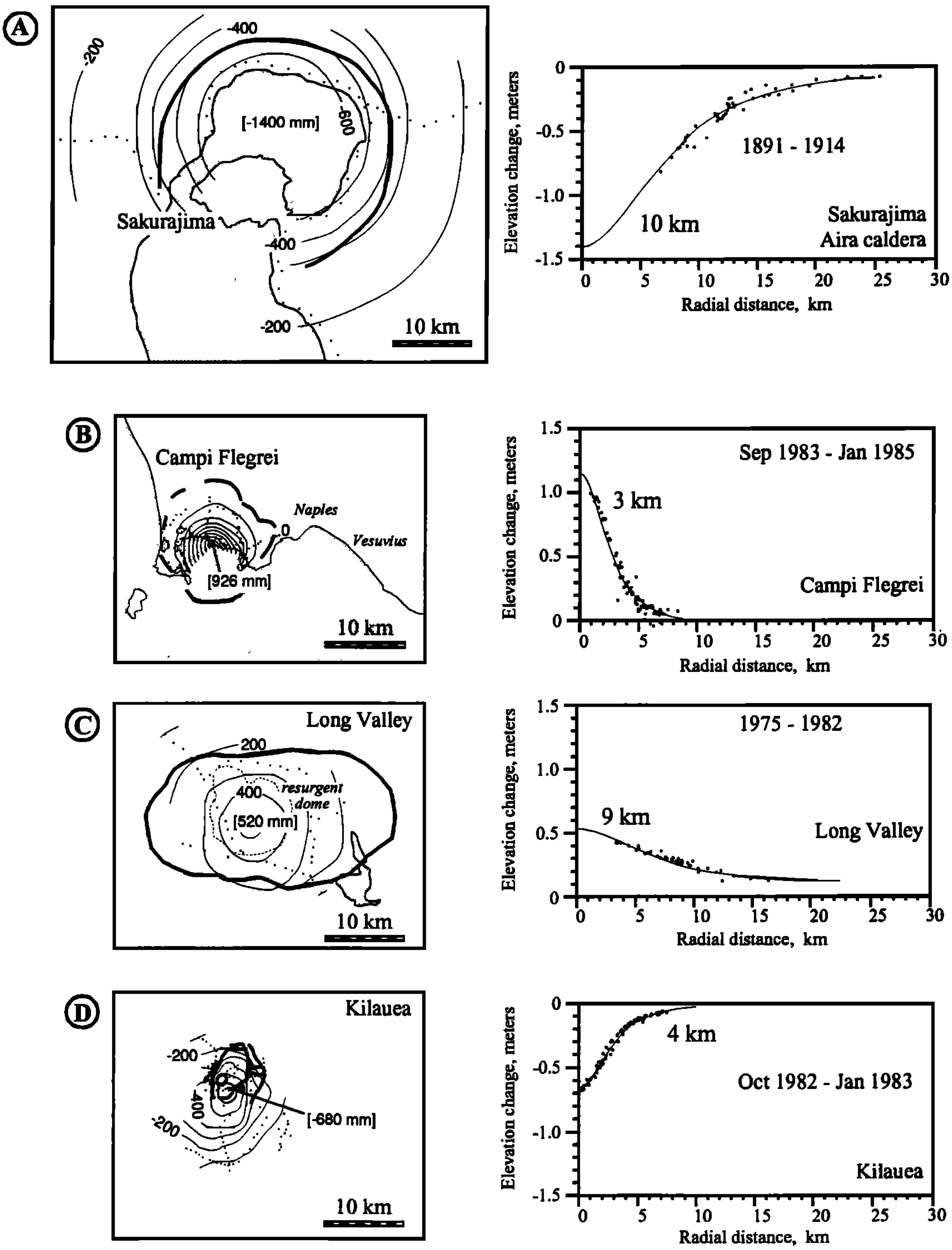
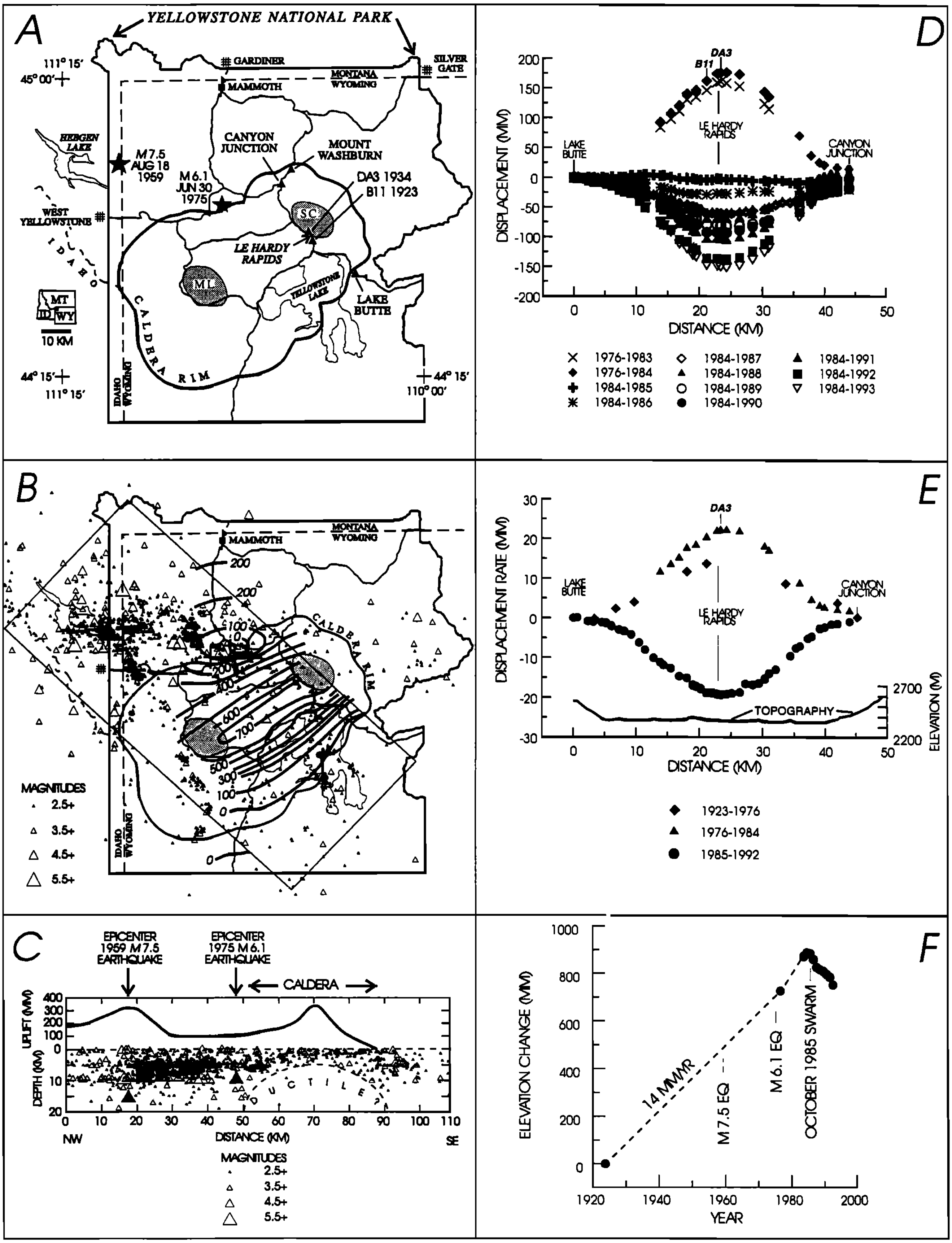


Figure 3. Contours of vertical displacement for the same four volcanic systems shown in Figure 2. The amount of inferred maximum vertical displacement is given in brackets for each system. Contour interval of elevation changes is 200 mm. The plots on the right compare measured (dots) and computed (solid line) elevation changes. Computed changes are based on a point model for a pressure source, parameterized by the depth to the source [*Mogi*, 1958]. (a) Sakurajima. From the 10-km-deep point model, as much as 1400 mm of subsidence probably occurred in the center of Aira caldera between 1891 and 1914 [*Hashimoto and Tada*, 1989]. (b) Campi Flegrei. A much shallower source is indicated, at a depth of 3 km, during a period of continuous uplift [*Dvorak and Berrino*, 1991]. (c) Long Valley. Since 1975, most of the vertical displacement can be reproduced by a single point source located at a depth of 9 km, so that there may have been as much as 520 mm of uplift by 1982 [*Savage et al.*, 1987]. (d) Kilauea. A depth of 4 km to a point source is indicated for the 5-day-long summit subsidence that accompanied the January 3, 1983, eruption along the east rift zone.


A
YELLOWSTONE NATIONAL PARK
GARDINER
SILVER GATE
MAMMOTH
MONTANA
WYOMING
HEBGEN LAKE
M7.5 AUG 18 1959
M6.1 JUN 30 1975
CANYON JUNCTION
MOUNT WASHBURN
DA3 1934
B11 1923
WEST YELLOWSTONE
IDAHO
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DISPLACEMENT (MM)
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C
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EPICENTER 1975 M6.1 EARTHQUAKE
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UPLIFT (MM)
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F
ELEVATION CHANGE (MM)
YEAR
14 MM/YR
M7.5 EQ
M6.1 EQ
OCTOBER 1985 SWARM

1990]. A similar change from slow uplift to slow subsidence, both movements at an average rate of about 20 mm yr^{-1}, has been measured at Askja caldera, Iceland, since 1966 [*Tryggvason*, 1989; *Rymer and Tryggvason*, 1993]. Except for Usu, none of these volcanoes has erupted since displacement measurements were begun.

The uplift profile across Yellowstone is essentially a mirror image of the subsidence profile [*Dzurisin et al.*, 1990] (Figure 4). The same was noted for Campi Flegrei [*Lirer et al.*, 1987]. This means that the location of the source that gave rise to uplift is near the location of the source that caused subsidence, even though the rates of uplift and subsidence were very different at Campi Flegrei (Figure 2b).

Kilauea. The fourth volcano with a long history of displacement measurements is Kilauea, Hawaii, a basaltic shield that is supplied magma from a mantle plume. Accurate leveling surveys from a distant benchmark to the summit area were begun in 1919–1920 [*Wilson*, 1935]. Frequent surveys of the summit area have been made since 1966 [*Fiske and Kinoshita*, 1969]. The summit area has subsided almost 5 m during the past 85 years and is now at its lowest elevation since leveling surveys were begun (Figure 2d). During this same period, Kilauea has erupted a few dozen times, mostly since 1952 [*Decker*, 1987].

Frequent activity at Kilauea has made this volcano a natural laboratory to develop and test techniques to record surface displacements. A useful technique has been the daily measurement of tilt made in an underground vault within 3 km of the point of maximum vertical movement. Between 1913 and 1961, such measurements were made using a mechanical seismograph [*Jaggar and Finch*, 1929]. Since 1956, more precise measurements have been made using a water tube instrument [*Eaton*, 1959]. In Figure 2d the daily tilt measurements are superimposed on the history of summit elevation changes. Though the tilt record of the mechanical seismograph is more erratic than the water tube instrument, the period of overlap from 1956 to 1961 shows a similar response of each instrument. Because Kilauea has been very active, superposition of tilt and elevation changes makes it possible to identify from the tilt records which sudden volcanic events caused the elevation changes.

The largest summit subsidence occurred from April to June 1924, when magma moved from a shallow summit reservoir to the east rift zone [*Jaggar and Finch*, 1929; *Dvorak*, 1992]. During the subsidence, steam explosions occurred from Halemaumau, a summit pit crater that enlarged during the 1924 activity. After 1924, summit tilt changes were slow and unrelated to a few small summit eruptions or to a shallow earthquake swarm in May 1938. In the spring of 1950, an increase in the rate of shallow earthquakes was concurrent with uplift of the summit region [*Finch and Macdonald*, 1953]. Since 1950 a now familiar displacement pattern has recurred in the summit area. During a period between successive rift zone eruptions, which may last from a few weeks to a few years, the surface rises at tilt rates of 1 μrad or more per day. For the instrument sites used at Kilauea, these tilt rates correspond to uplift of as much as several millimeters per day. Eventually, the uplift is interrupted by a sudden subsidence, which may coincide with a rift zone intrusion and, possibly, eruption. Within a few hours the caldera floor may drop several tens to hundreds of millimeters.

An almost identical and recurring pattern of relatively slow uplift at a rate of a few millimeters per day, interrupted by sudden subsidence and a rift zone intrusion or eruption, began in 1975 at Krafla volcano, Iceland [*Björnsson*, 1985]. Until 1975, Krafla had been dormant since the previous series of eruptions that occurred from 1724 to 1729. A comparison of triangulation surveys made in 1938 and 1965 showed no significant displacements [*Björnsson et al.*, 1979]. The next survey in 1971 suggested slight contraction. Increased seismic activity around Krafla in early 1975 prompted another survey, which was conducted a month or so before an eruption in December 1975. The results showed significant expan-

Figure 4. (opposite) The Yellowstone region, including Yellowstone caldera and an adjacent fault zone. (a) Outline of Yellowstone caldera, Mallard Lake resurgent dome (ML), and Sour Creek resurgent dome (SC), plus locations and features mentioned in the text. Thin lines that connect Mammoth, Canyon Junction, Silver Gate, and West Yellowstone are routes for repeated leveling surveys. The entire network was measured in 1923, 1975–1976, and 1988; the traverse between Canyon Junction and Lake Butte was measured each year from 1983 to 1993. Also shown are the locations of two large earthquakes (stars) and two bench marks (DA3 1934 and B11 1923) near the center of uplift and subsidence at Le Hardy Rapids. (b) Locations of $M > 2.5$ earthquakes for the period 1959–1989 (triangles) and contours of crustal uplift (heavy lines, labeled in millimeters) for the period 1923–1976. The rectangle encloses earthquakes shown in cross section in Figure 4c. (c) Northwest-southeast cross section showing a generalized uplift profile for the period 1959–1993 and hypocenters of earthquakes outlined in Figure 4b. Note the aseismic region beneath the caldera, where high temperatures result in ductile, rather than brittle, behavior of the crust. The uplift profile does not include coseismic movements associated with the $M = 7.5$ Hebgen Lake earthquake in August 1959. (d) Uplift and subsidence profiles between Lake Butte and Canyon Junction measured in 1976 and each year from 1983 to 1993. Le Hardy Rapids rose at an average rate of 22 mm yr^{-1} during 1976–1984, stopped moving during 1984–1985, and started to subside during 1985–1986. Since then the subsidence rate has varied from about 10 mm yr^{-1} to 30 mm yr^{-1}. (e) Average uplift and subsidence rates between Lake Butte and Canyon Junction for three intervals spanned by leveling surveys. The uplift and subsidence profiles are virtually mirror images of each other, indicating that the locations of the sources are likewise similar. (f) Elevation as a function of time at bench mark B11 near Le Hardy Rapids as measured by repeated leveling surveys. Note the large time intervals between surveys before 1983.

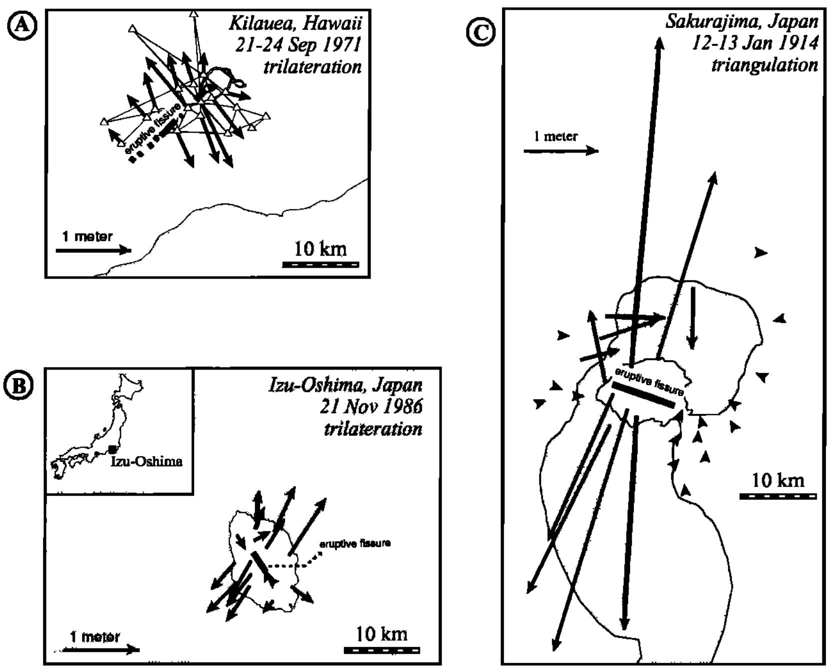


Figure 5. Horizontal displacements associated with three fissure eruptions. (a) The September 21–24, 1971, eruption along the southwest rift zone of Kilauea [*Duffield et al.*, 1982]. The trilateration network is shown as thin lines that connect station pairs; each station is represented as a small triangle. (b) The November 21, 1986, eruption of Izu-Oshima [*Hashimoto and Tada*, 1990]. (c) The January 12–13, 1914, eruption of Sakurajima, on the southern rim of Aira caldera. These displacements were determined from triangulation surveys conducted in 1895–1898 and 1914 [*Hashimoto and Tada*, 1989].

sion. During the December 1975 eruption the surface of Krafla subsided as much as 2.5 m [*Björnsson*, 1985]. During the next half year, the surface rose at an average rate of 6 mm d^{-1}, then subsided suddenly in September as magma moved from beneath the summit area northward along a subterranean conduit system. During the next 10 years the displacement pattern alternated between relatively slow uplift and rapid subsidence. Many episodes of subsidence coincided with rift zone eruptions.

A wide range of displacement rates has been recorded at the few volcanoes where measurements span several decades. The range of rates suggests that different processes are important, not solely the sudden supply of magma to feed an imminent eruption. Moreover, a change of displacement rate might signal a change in the dominant process causing surface displacement. In a later section we use displacement rate as a means of distinguishing among various processes that cause volcanoes to deform. For the remainder of this section we concentrate on the largest and most rapid displacements recorded at active volcanoes: displacements related to development of fissures and central vents.

Development of Fissures

A basaltic eruption often begins from a long fissure, which may be discontinuous and extend several hundred meters to a few tens of kilometers in length. The elongated geometry of a fissure suggests that the magma conduit feeding such an eruption is a dike that wedged open the upper crust and reached the surface.

Trilateration or triangulation measurements across basaltic fissures show that the surface opens perpendicular to the axis of a fissure. For example, horizontal displacements derived from trilateration measurements made across the southwest rift zone of Kilauea showed about 2 m of opening across the axis of the rift zone during the September 1971 eruption [*Duffield et al.*, 1982] (Figure 5a). This amount of opening is within the range of dike widths measured at eroded Hawaiian volcanoes [*Walker*, 1987]. Trilateration measurements made at other basaltic volcanoes show a similar pattern of crustal opening across fissures, for example at Izu-Oshima, Japan [*Hashimoto and Tada*, 1990] (Figure 5b); Etna, Italy [*Murray and Pullen*, 1984]; and Krafla, Iceland [*Sigurdsson*, 1980]. The amount of opening measured during the 1914 eruption of Sakurajima is the largest yet recorded, an average of 8 m along a 7-km-long fissure [*Hashimoto and Tada*, 1989] (Figure 5c).

Though the maximum horizontal displacement during the 1914 eruption was associated with formation of the eruptive fissure that formed across Sakurajima, the maximum vertical displacement occurred over the center of Aira caldera (compare Figures 3a and 5c). Evidently, a reservoir beneath the caldera supplied magma that moved laterally about 10 km and, perhaps, was erupted. Such lateral movement of magma from a central reservoir is common, especially at basaltic volcanoes. Subsidence of the summit area and eruption along a flank have been concurrent at Mauna Loa, Hawaii [*Lockwood et al.*, 1987]; at Piton de la Fournaise, Réunion [*Delorme et al.*, 1989]; and at Tolbachik, Kamchatka [*Fedotov et al.*, 1980]. The largest distance recorded between a flank eruption and concurrent summit subsidence was 50 km during the 1975 eruption of Krafla [*Björnsson et al.*, 1979] and during the 1960 eruption of Kilauea [*Eaton and Murata*, 1960].

The pattern of vertical displacement associated with a fissure eruption is uplift on either side of the fissure and subsidence along the axis of the fissure, sometimes called a ridge-trough-ridge structure [*Pollard et al.*, 1983]. The surface on either side of the fissure rises in order to accommodate the volume of the intrusion. Directly over the fissure the sideways opening of the intrusion stretches the overlying rock and causes the surface to drop. If the amount of dilation is large enough to crack the surface, then normal faults develop on either side of the fissure and the central graben may drop a meter or more. This pattern has been measured at several basaltic volcanoes, notably Kilauea (Figure 6), Etna (Figure 7), and Krafla [*Sigurdsson*, 1980]. Also, it occurred along the short segment where the mid-ocean ridge system comes onto land in Asal-Ghoubbet, Djibouti [*Ruegg et al.*, 1979] (Figure 8).

Frequent geodetic measurements, such as from continuously recording tiltmeters, show that fissures often grow almost to their final size in a day or less. Kilometer-long eruptive fissures formed within a few hours or less in the summit areas of Kilauea in April and September 1982 (Hawaiian Volcano Observatory, unpublished data, 1982) and of Piton de la Fournaise in 1983 and 1984 [*Zlotnicki et al.*, 1990]. Longer fissures that form on volcano flanks reach most of their final lengths of 5–10 km and maximum openings of 1–3 m within a day, for example at Kilauea in 1983 [*Okamura et al.*, 1988], Krafla in September 1977 [*Hauksson*, 1983], and Etna in 1989 [*Bonaccorso and Davis*, 1993]. If an eruption persists a week or more, a fissure may extend a few more kilometers in length and open an additional few tenths of a meter [*Ferrucci et al.*, 1993; *Dvorak et al.*, 1986].

One of the most complete recordings of the growth of an eruptive fissure was made in 1989 before an undersea eruption off the east coast of Izu Peninsula, Japan [*Yamamoto et al.*, 1991] (Figures 9a and 9b). *Okada and Yamamoto* [1991] identified five phases of the development of the fissure during the first 2 weeks of July 1989. These phases were based on changes in displacement rate measured by a tiltmeter and by a dilatometer. The latter records volumetric strain changes. In addition to the automatic instruments, daily measurements were made of the relative position of two bench marks located across the eventual eruptive fissure using trilateration and the satellite-based GPS. These different techniques showed consistent displacements and strain changes during an intense shallow earthquake swarm that lasted 1 week (Figure 9c).

Development of Central Vents

Compared with the broad displacement patterns recorded at explosive calderas and basaltic volcanoes, patterns at stratovolcanoes are localized. At Mount St. Helens, displacements related to rock failure were recorded only over an area of 4 km^2 on the north flank during the several weeks that preceded structural failure of the flank and the beginning of a Plinian eruption in 1980 [*Lipman et al.*, 1981]. The amount of horizontal displacement of the north flank, however, was several hundred meters prior to the eruption. Large, localized failure of the volcano occurred also before the 1956 Plinian eruption of Bezymianny, Kamchatka [*Gorshkov*, 1959]. Within explosive calderas, displacements may be large and localized and last a few years, such as those at Usu in 1910–1911, 1943–1945, and 1977–1982 [*Omori*, 1913; *Minakami et al.*, 1951; *Yokoyama et al.*, 1981]. The large displacements at Mount St. Helens, Bezymianny, and Usu were caused by dacitic magma intruded within a few hundred meters of the surface, forming a cryptodome. At Usu, slight displacements of a few tens of millimeters were measured as far as 8 km from the uplifted dome. Comparable amounts of displacement have not been recorded several kilometers from the few stratovolcanoes where geodetic measurements have been made.

The lack of measured displacements around stratovolcanoes may be due in part to the difficulty of making such measurements on the steep and unstable slopes that compose these volcanoes. Where intensive measurements have been made, however, the amount of displacement has been small. A decade of annual tilt measurements at Soufrière of St. Vincent, Lesser Antilles, spanned a period of vertical explosions, eruption of pyroclastic flows and dome growth [*Shepherd et al.*, 1979]. Even though a major eruption occurred, tilt changes were 20 μrad or less at distances of 2–6 km from an eruptive vent in the summit crater [*Fiske and Shepherd*, 1990]. At most, the flanks of Mount St. Helens moved outward a few tens of millimeters before major magmatic explosions in July and August 1980 [*Swanson et al.*, 1981]. Before episodes of nonexplosive dome growth within the crater, no displacements were measured on the outer flanks; however, displacements of a meter or more were measured within the crater across thrust faults [*Swanson et al.*, 1983].

The example of Mount St. Helens stresses the need to make measurements close to the central vent. This is also borne out at other stratovolcanoes. Annual leveling surveys begun in 1967 along a 1-km-long line at White Island, New Zealand, revealed relative displacements that were less than 10 mm at bench marks farther than 100 m from the explosion craters [*Clark*, 1982; *Clark and Otway*, 1989]. Within 100 m, elevation changes were 100 mm or more. Trilateration measurements across the summit crater of Ruapehu, New Zealand, showed that the summit extended only several tens of millimeters before steam explosions, even when measurements were made a few minutes before an explosion [*Otway*, 1979]. Unless the volcano slope fails and a major bulge forms, as at Mount St. Helens or Usu, measurements made more than 1 km from the vent must have a precision of 1 cm or less in order to detect displacements associated with eruptions of most stratovolcanoes.

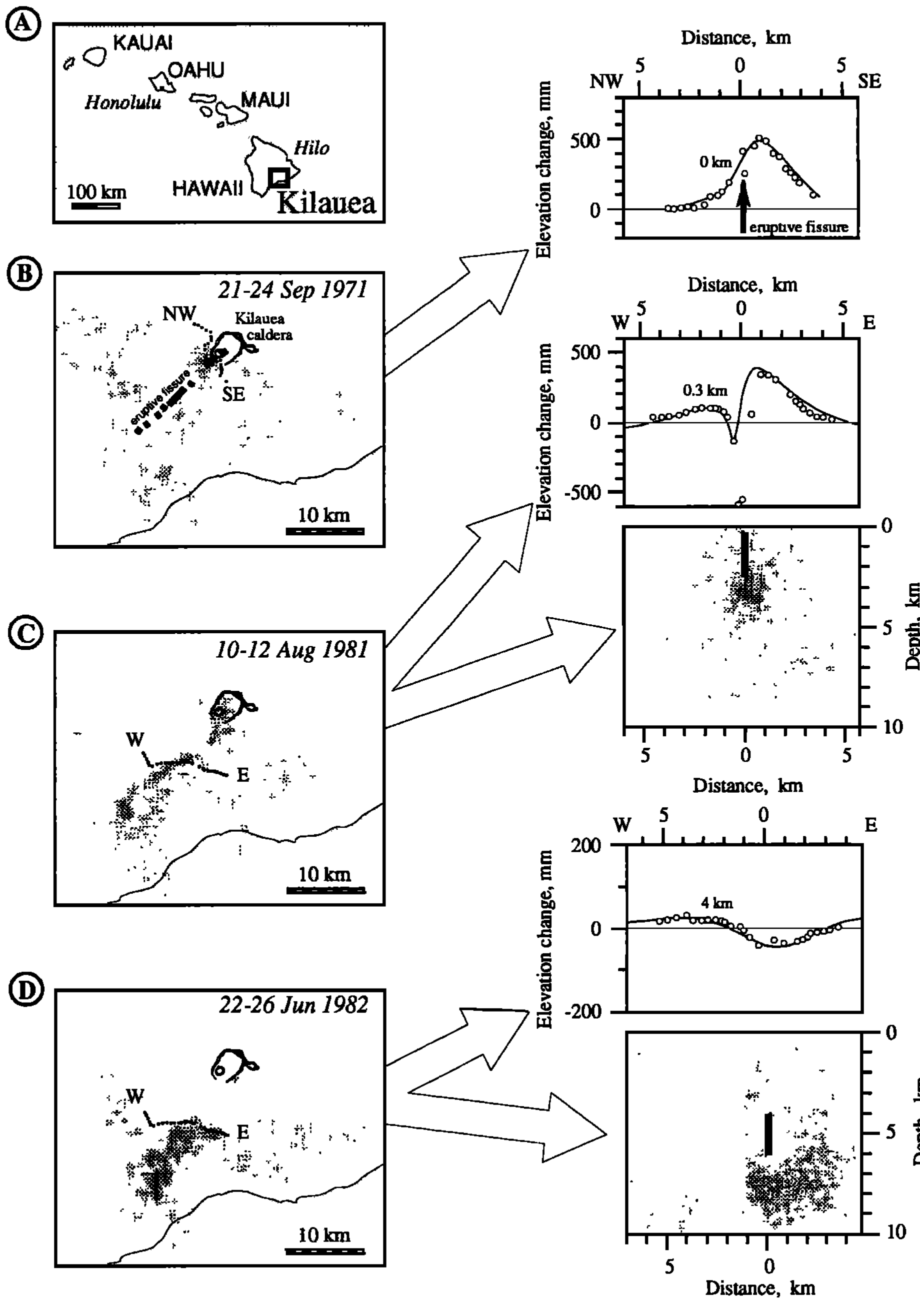


Figure 6. Three rifting events along the southwest rift zone of Kilauea. (a) Location map showing the major Hawaiian Islands and the location of Kilauea. (b) (left) Epicenters of earthquakes (plus symbols) recorded during the September 21–24, 1971, eruption along the southwest rift zone. The extent of the eruptive fissure is indicated as a thick, segmented line [*Duffield et al.*, 1982]. Locations of bench marks along a short leveling route are shown as small dots that cross the southern region of the summit caldera. (right) Measured (circles) and computed (solid line) elevation changes along the leveling route shown to the left [*Dvorak*, 1990]. (c) (left) Epicenters of earthquakes recorded during the August 10–12, 1981, intrusion into the southwest rift zone. Locations of bench marks along a leveling route that crosses the southwest rift zone are shown as small dots. (right) Measured and computed elevation changes and earthquake hypocenters projected onto a vertical plane perpendicular to the rift axis. The distance between the two maxima in the elevation profile suggests that the dike rose to within a few hundred meters of the surface. The thick line segment, superposed on earthquake hypocenters, shows the vertical extent of the pressurized crack used to compute elevation changes [*Pollard et al.*, 1983]. (d) (left) Epicenters of earthquakes recorded during the June 22–26, 1982, intrusion into the southwest rift zone. The leveling route is the same as that in Figure 6c. (right) Measured and computed elevation changes and earthquake hypocenters projected onto a vertical plane perpendicular to the rift axis. Compared with the results in Figure 6c, the greater distance between maxima in the elevation changes for June 1982 suggests a deeper source [*Dvorak et al.*, 1986].

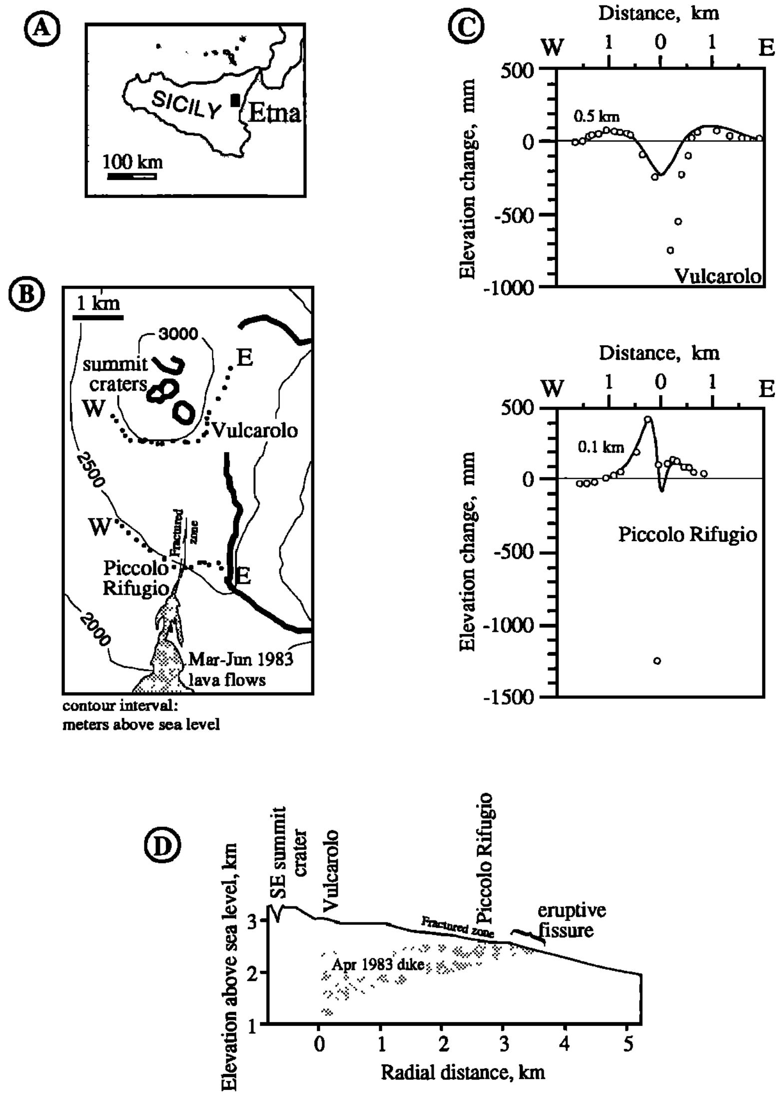


Figure 7. (a) Location map of Etna, Sicily, Italy. (b) Location of two leveling routes, one near the summit (Vulcarolo) and one near the 2500-m elevation (Piccolo Rifugio). The lava flows of March–June 1983 are indicated by a stippled pattern. A fractured zone formed between the summit and the March 1983 eruptive vent. (c) Measured (circles) and computed (solid line) elevation changes along the two leveling routes located in Figure 7b. The depth to the top of the dike is given in each plot. (d) Cross section of the dike that grew horizontally from a column of magma beneath the summit area and intersected the south flank of Etna just below the 2500-m elevation. (Modified from *Murray and Pullen* [1984, Figures 1–4]; copyright Springer-Verlag.])

VISUALIZING THE SOURCE OF GROUND MOVEMENT

The simple displacement patterns shown in Figures 3–8 might seem to be at odds with the internal complexity evident at eroded volcanoes. The many intrusive bodies exposed at an eroded volcano, however, are the culmination of many hundreds, or even thousands, of events, whereas the displacement patterns reflect one event or a few events. Moreover, even though an individual dike or sill may show abrupt irregularities, the scale of an irregularity is usually small in comparison with the dimensions of a dike or sill. Therefore, as was pointed out by *Pollard et al.* [1983], a mechanical interpretation of displacement patterns in terms of buried sources can make use of Saint-Venant's principle. The principle states that the irregularities of a pressure source have a negligible effect on the resultant displacement pattern at distances large compared with the scale of the irregularities [*Love*, 1944]. In essence, rocks surrounding a source act as a low-pass filter on the details of the source. Thus surface displacement patterns can be interpreted in terms of simple source geometries that represent sills, dikes, or plugs. The derived dimensions and subsurface displacements are average values of the source.

A second concern is whether it is valid to assume that the mechanical behavior of a volcano is elastic. Open cracks, fault scarps, layers of erupted deposits, and heated ground would seem to invalidate the assumption. The effect of a crack or discontinuity alters the elastic constants but does not change the fundamental elastic behavior [*Jaeger and Cook*, 1977]. The creation of a new crack, such as the formation of a dike, is enigmatic in that a new discontinuity is introduced on which the shear stress vanishes and hence would be expected to produce large changes in the stress field [*McKenzie et al.*, 1992]. Observations show that this is not the case. For example, dike patterns that represent dozens of events, such as Spanish Peaks, Colorado, can be described by elastic theory, even though multiple dikes formed [*Muller and Pollard*, 1977]. Because the theory provides a good basis

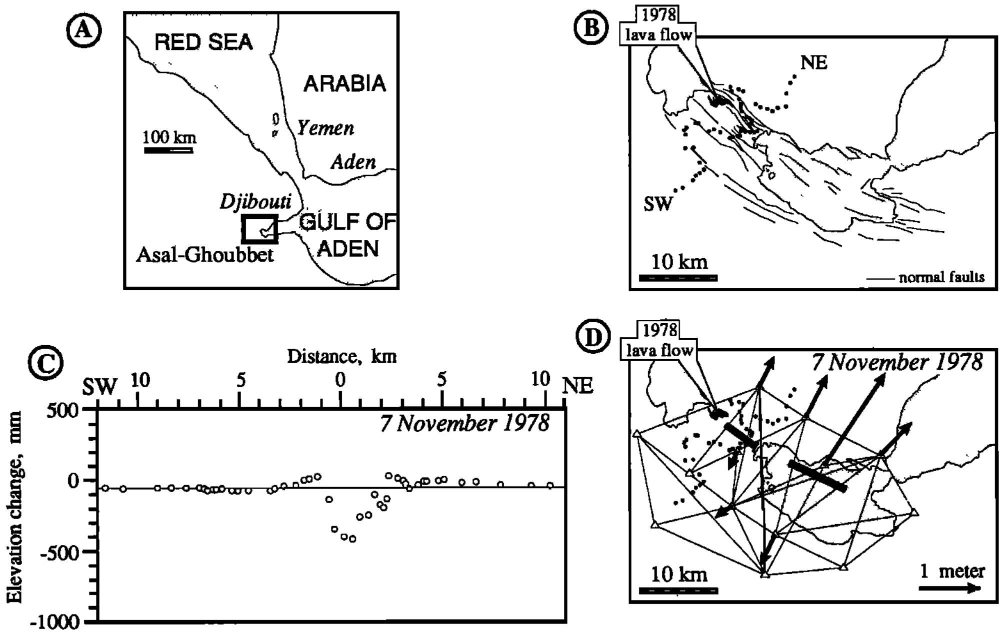


Figure 8. (a) Location map of Asal-Ghoubbet, Djibouti. (b) Location of a leveling route that crosses the East Africa rift system (dots). Normal faults, indicated as thin lines, are aligned along the axis of the rift system. A small lava flow, shown as a black patch, erupted during extension of the rift system on November 7, 1978. (c) Elevation changes along the leveling route shown in Figure 8b associated with the 1978 eruption and rift extension. (d) Horizontal displacements associated with the 1978 eruption and rift extension. Displacements are shown as vectors; trilateration stations as small triangles. The trilateration network is shown by thin lines that connect station pairs. The two thick line segments are the location and horizontal extent of pressurized cracks that can reproduce the measured horizontal displacement. (Modified from *Ruegg et al.* [1979].)

to describe many observations, it is used here to understand the causes of surface displacement in terms of magma bodies of simple geometries. Though a component of displacement at many active volcanoes is probably anelastic, the limited coverage and precision of displacement measurements seldom warrant the use of an inelastic model.

Point Source of Dilation

A few field experiments have been conducted in which a slurry of material was injected forcefully into the ground, an experiment analogous to the forceful injection of magma within a volcano. During one such experiment, 2500 m^3 of a sand-water mixture was injected repeatedly at the bottom of a drill hole at depths of 265–290 m [*de Laguna et al.*, 1968]. After the first two injections, drilling at a dozen closely spaced holes showed that the sand-water mixture formed two 10-mm-thick sills that extended as much as 30 m outward from the injection points. Measurements of surface displacement showed that the injections raised the surface 20 mm near the drill hole, and the amount of displacement decreased smoothly with distance from the drill hole.

The uplift pattern that resulted from this experiment can be modeled as the dilation of a point source embedded in a homogeneous, isotropic, elastic half-space. This model represents pressurization of a spherical cavity in which the cavity diameter is much smaller than the depth of burial. Simple analytic equations of surface displacement caused by dilation of a point source were published by *Mogi* [1958] and *Okada* [1985]. (An extended source that represents the sill could also reproduce the uplift pattern, though the measurements did not require this more complicated model.) The resultant uplift pattern of a point source is a simple, bell-shaped curve that is characterized by two parameters: (1) depth of the source and (2) volume of uplift. In the few injection experiments that have been reported, the volume of uplift was within 20% of the volume of injected material [*Davis*, 1983]. Within the uncertainty of computing an uplift volume and the volume change caused by possible compaction of the sand-water mixture, the injected and uplifted volumes were equal. For the experiments described above, the uplift pattern was reproduced by a dilatational point source at a depth of 270 m, within the depth range of the injection experiments.

An unintended field experiment that was converse to the injection experiments has also been reported: measurement of surface displacements caused by 35 years of pumping of the Lacq gas field, France [*Maury et al.*, 1992]. The gas reservoir has a thickness of about 500 m, and its top is located 3 km below the surface. In these withdrawal experiments the subsidence pattern was a simple, inverted bell shape that could be reproduced as a contracting point source at a depth of 3 km.

The relation between these field experiments and

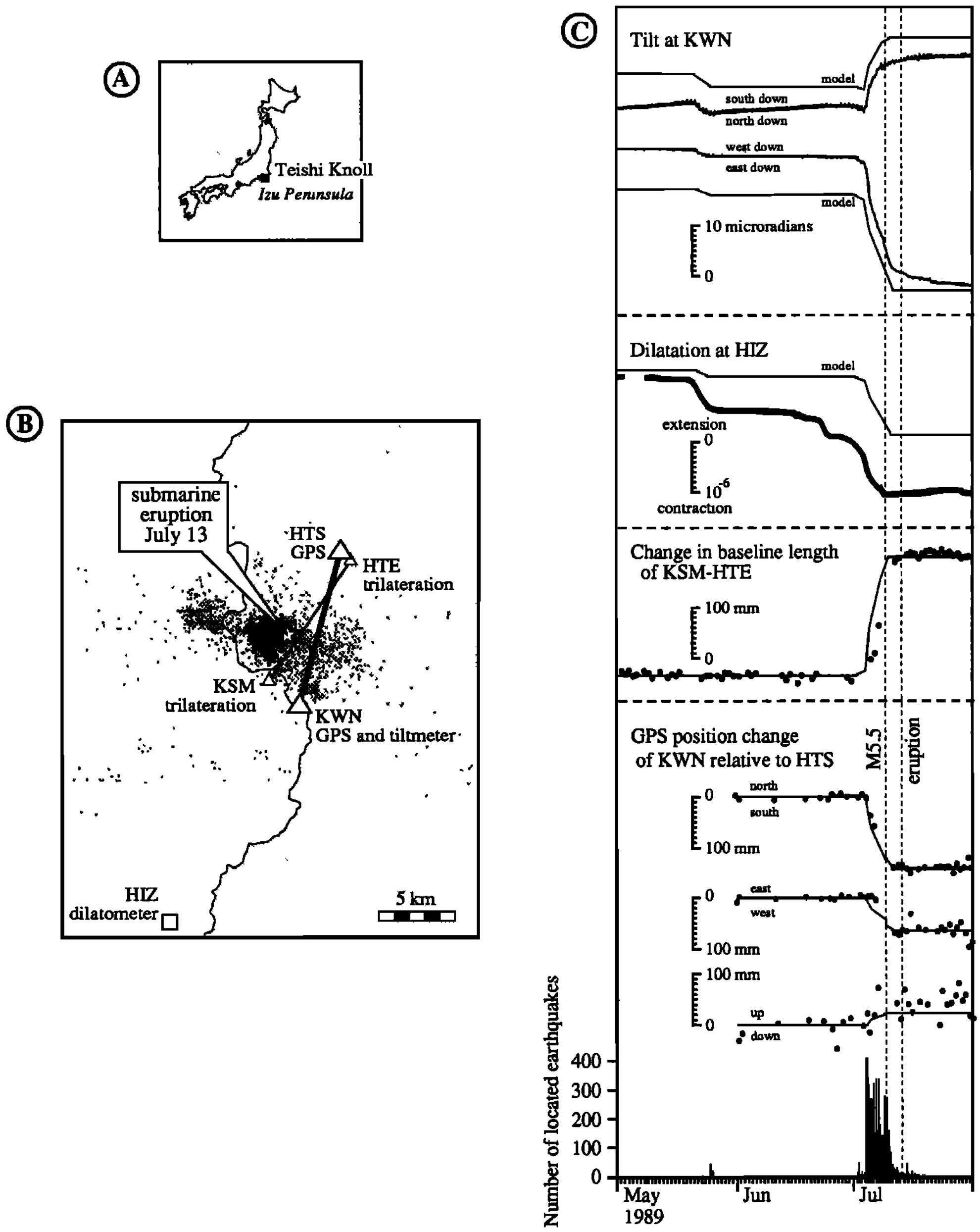


Figure 9. (a) Location of Izu Peninsula, Japan. (b) Teishi Knoll, an undersea volcano located a few kilometers off the east coast of Izu Peninsula. Earthquake epicenters recorded from May through July 1989 are shown as small dots. The locations of different geodetic stations and instruments are indicated by the square and triangles. (c) May–July record of geodetic changes and earthquake counts. The records of the two-component tiltmeter at KWN were sensitive enough to record earth tides. Model calculations, based on two vertically growing dikes and strike-slip displacement of a $M = 5.5$ earthquake, are indicated by the two smooth lines. The record of the dilatometer at HIZ is indicated by a thick line; model calculations are indicated by the thin line. The net change in baseline length, determined by frequent trilateration measurements between KSM and HTE, was about 200 mm of extension. Frequent GPS measurements revealed the three-dimensional relative movement of KWN relative to HTS. Hourly earthquake counts are at the bottom. Note that no major ground displacement or earthquake swarm was coincident with the explosive submarine activity on July 13. (Modified from *Okada and Yamamoto* [1991].)

active volcanoes is obvious. The injection experiments are similar to intrusion of a small magma body; the pumping of a gas field has characteristics that are similar to withdrawal of magma from a shallow reservoir. Though the injected body and the gas reservoir are not point sources, such a model did locate the centroid of the actual source and, at least for the injection experiments, show an equivalence between injected volume and uplifted volume.

As applied to active volcanoes, several alterations have been made to the dilatational point source described by *Mogi* [1958] so that the model is a more realistic approximation of a pressurized magma body. *Davis* [1986] derived equations of surface displacement for a pressurized ellipsoidal cavity. He suggested that the pattern of vertical and horizontal displacements of Kilauea was better explained by pressurization of a vertically elongated ellipsoid than by pressurization of a sphere. *Dieterich and Decker* [1975] used a finite-element method and assumed homogeneous elasticity throughout a half-space to compute surface displacements of axisymmetric magma bodies represented by a sphere, ellipsoid, or thin disk. The various assumed geometries were not distinguished by the patterns of vertical displacement; horizontal displacements were needed to determine the basic geometry of a body. *Bianchi et al.* [1987] used a finite-element method to determine the effect of inhomogeneous elasticity that surrounds the pressurization of a proposed magma body beneath Campi Flegrei. The inhomogeneities were assumed to arise either from heating of rock by the magma body or from extensive fractures created by intense earthquake swarms above the magma body. *Bianchi et al.* [1987] showed that reasonable values for magma pressure could produce the few meters of uplift recorded recently at Campi Flegrei.

Though these and other proposed alterations to the point source model are more realistic, they introduce more complexity and hence more model parameters. The limited extent of geodetic measurements available for most volcanoes cannot distinguish between a point source of dilation and a more complex model for a shallow magma reservoir. Even for the most intensively studied volcanoes, such as Kilauea, Long Valley, Campi Flegrei, and Sakurajima, geodetic measurements are seldom sufficiently dense and diverse to distinguish between a point source model and a more complex model. For these reasons we consider only point source models in the remainder of this subsection. Later, we consider other geometries when a measured pattern of surface displacement clearly indicates an extended body, such as a dike or sill.

The point source model successfully reproduces the patterns of vertical displacement at many volcanoes during either uplift or subsidence, for example, the four volcanoes in Figure 3. In addition, the model reproduces tilt patterns at Izu-Oshima [*Oikawa et al.*, 1991], Piton de la Fournaise [*Lenat et al.*, 1989b], and Tangkuban Parahu, Indonesia [*Dvorak et al.*, 1990] and patterns of horizontal strain at Rabaul [*Archbold et al.*, 1988], Long Valley [*Denlinger and Riley*, 1984], and Kilauea [*Dvorak et al.*, 1983]. For the latter two volcanoes an additional source of horizontal strain is needed. At Long Valley the addition is displacement of a normal fault near the south caldera rim. At Kilauea, permanent north-south extension is caused by emplacement of elongated dikes that strike east-west and seaward movement of the south flank. In the few cases where tilt or trilateration surveys were conducted at the same time as leveling surveys, the same depth was computed for a point source model regardless of which type of survey results were used (for Kilauea [*Dvorak et al.*, 1983], Krafla [*Tryggvason*, 1986] and Mauna Loa [*Decker et al.*, 1983]).

Depth of a dilatational point source has been determined at more than a dozen volcanoes (Table 2). Some shallow sources, less than about 2 km below the surface, might be related to pressure increase or release of a hydrothermal system, such as at Tangkuban Parahu, Indonesia [*Dvorak et al.*, 1990]. At a volcano where magmatic eruptions are frequent, such as Piton de la Fournaise [*Lenat et al.*, 1989a, b], Etna [*Murray and Guest*, 1982], and White Island [*Clark and Otway*, 1989], a shallow point source is attributed to intrusion of a small magma body that later fed an eruption.

At basaltic shield volcanoes in Hawaii and Iceland, where frequent leveling surveys have allowed the determination of dozens of point source solutions, most sources lie at a depth of about 3 km. At these volcanoes this depth corresponds to the level of neutral buoyancy, where the density of basaltic magma equals the density of the surrounding rock [*Ryan*, 1987]. If magma rises through the crust by buoyancy, then it should accumulate at this level. Further rise requires pressurization of a magma body in excess of lithostatic pressure or a decrease in density of the magma body, perhaps by exsolution of dissolved gases and retention of the resulting bubbles within the magma.

Often the position of the point source is at the top of an aseismic region that is capped and partially surrounded by earthquakes. Notable examples are Kilauea [*Tilling and Dvorak*, 1993], Mauna Loa [*Decker et al.*, 1983], Campi Flegrei [*Dvorak and Berrino*, 1991], and Izu-Oshima [*Hashimoto and Tada*, 1990]. The aseismic region may coincide with a region of low seismic velocity or, in the case of Campi Flegrei, be defined by an abrupt change of velocity structure. The coincidence at the same zone within a volcanic system of low velocity, lack of seismicity, and location of a pressure source is strong evidence of magma storage.

Reliable point source solutions deeper than 6 km have been determined only at large calderas. In these cases the source lies beneath the caldera center (Table 2). These sources are capped and partially surrounded by regions of intense seismicity and are located within resurgent calderas, such as Yellowstone (Figure 4) and Long Valley [*Langbein et al.*, 1993]. At these two calderas the deep point source lies within the upper portion of a low-velocity region interpreted as a partially molten pluton [*Lehman et al.*, 1982; *Steeples and Iyer*, 1976]. A shallow point source located at depths of 3–6 km has also been estimated for Yellowstone [*Vasco et al.*, 1990].

At least two point sources, each representing a separate magma body, are needed to explain the 1975–1982 pattern of vertical displacements at Aira caldera [*Sakurajima Volcanological Observatory*, 1989]. One body

TABLE 2. Depth of Pressure Centers

Volcano	*Time Period*	*Depth,* km*	*Measurement*	*Reference*
Piton de la Fournaise, Réunion	Dec. 1983 to Jan. 1984	1.3	spirit level tilt	*Lenat et al.* [1989a, b]
Krafla, Iceland	March 1976 to July 1977	3	leveling	*Björnsson et al.* [1979]
	1975–1985	2–4	leveling	*Ewart et al.* [1991]
Askja, Iceland	1986–1987	2.5	leveling	*Tryggvason* [1989]
Izu-Oshima, Japan	1986–1987	4	tiltmeters	*Oikawa et al.* [1991]
Mauna Loa, Hawaii	1977–1981	3	leveling	*Decker et al.* [1983]
Tangkuban Parahu, Indonesia	1981–1986	1.5	spirit level tilt	*Dvorak et al.* [1990]
Etna, Italy	July 1975 to Sept. 1977	1.6	leveling	*Murray and Guest* [1982]
Campi Flegrei, Italy	Jan. 1982 to June 1984	3	leveling	*Berrino et al.* [1984]
Kilauea, Hawaii	1920–1927	3	leveling	*Mogi* [1958]
	1966–1967	2–4	leveling	*Fiske and Kinoshita* [1969]
	1959–1983	2–6	tilt and leveling	*Tilling and Dvorak* [1993]
Miyakejima, Japan	1982–1986	5†	leveling	*Tada and Nakamura* [1988]
Klyuchevskoy, Kamchatka	1981–1988	[3–10]	leveling	*Fedotov et al.* [1990]
White Island, New Zealand	1973–1976	0.5	leveling	*Clark and Otway* [1989]
Aira and Sakurajima, Japan	1891–1932	10	leveling	*Mogi* [1958]
	1891–1914	10	leveling	*Yokoyama* [1986]
	1975–1982	4 and 10	leveling	*Sakurajima Volcanological Observatory* [1989]
Rabaul, Papua New Guinea	1981–1982	1–2	spirit level tilt	*McKee et al.* [1985]
		1.2	tilt and leveling	*Archbold et al.* [1988]
Vulcano, Italy	March 1980 to Sept. 1981	[6.5]	leveling	*Ferri et al.* [1988]
Medicine Lake, California	1954–1989	9	leveling	*Dzurisin et al.* [1991]
Yellowstone, Wyoming	1923–1976	3–6	leveling	*Vasco et al.* [1990]
	1985–1987	10	leveling	*Dzurisin et al.* [1990]
Long Valley, California	1975–1980	11	leveling	*Savage and Clark* [1982]
	1980–1983	5 and 8	leveling	*Rundle and Whitcomb* [1984]
	1982–1983	13	leveling	*Denlinger et al.* [1985]
	1989–1991	7	trilateration	*Langbein et al.* [1993]
Karymsky, Kamchatka	1979–1980	[1.5]	leveling	*Magus'kin et al.* [1982]
Colima, Mexico	1990–1991	2–4	leveling	*Murray* [1993]

*Depths given in brackets are uncertain because of limited coverage of geodetic network.
†*Tada and Nakamura* [1988] report 8 km; our reanalysis of only preeruption data suggests 5 km.

lies 10 km beneath the caldera center. The other body is 4 km beneath Sakurajima, the basaltic-andesitic volcano on the south caldera rim and the site of all historic eruptions. Presumably, a large magma body that lies beneath the caldera center supplies magma to the shallower body, and the shallower body supplies magma to Sakurajima. During the several months after the end of the 1914 eruption, magma continued to move upward and laterally from the deep body to the shallow body.

Multiple point sources that represent magma storage have been identified at other volcanoes. At Kilauea, point sources are used to infer magma storage beneath the summit area and along the east rift zone [*Tilling and Dvorak*, 1993]. In some cases, summit subsidence and uplift of the rift zone are concurrent, suggesting that magma from the summit reservoir moved laterally and filled a reservoir along the east rift zone [*Jackson et al.*, 1975]. At Long Valley the location of two point sources beneath the resurgent dome may represent shallow cupolas of a partially molten pluton [*Rundle and Whitcomb*, 1984; *Langbein et al.*, 1993].

Plane Strain

The bilateral symmetry of horizontal displacements associated with fissure eruptions (Figure 5) argues that the underlying source is greatly elongated and cannot be considered as a point source. Most likely, such a source is actually a dike, and so an appropriate mathematical model would be a vertical, thin pressurized sheet. Patterns of horizontal displacement, however, are usually too sparse to develop an intuitive feeling of how the pattern changes for different dike parameters, such as dike height or depth of burial. A better means is to consider profiles of vertical displacement across a fissure.

Except near the ends of a dike, such profiles approximate the condition of plane strain, so that displacements are only perpendicular, not parallel, to the fissure. Using that approximation, *Pollard et al.* [1983] showed that the profile of vertical displacement depends on three parameters: (1) height of the dike, (2) depth to the top of the dike, and (3) dip angle.

As an illustration, the profiles of three dikes intruded into the southwest rift zone of Kilauea are compared in Figure 6. The top profile is across an eruptive fissure: The dike reached the surface, and so a single maximum occurs in the profile. The lower two profiles are across dikes that did not reach the surface, and so each profile has two maxima separated by a trough. For the 1981

TABLE 3. Dimensions of Dikes

Volcano	*Date of Event*	*Depth of Top, km*	*Height, km*	*Length, km*	*Width, m*	*Reference*
Piton de la Fournaise,	Dec. 4, 1983, and	0.3	1.0	1.8	1.0	*Zlotnicki et al.* [1990]
Réunion	Jan. 18, 1984	0.3	1.0	2.2	1.0	
Krafla, Iceland	Dec. 20, 1975	···	···	···	1.5	*Sigurdsson* [1980]
	Jan. 7, 1978	0.5	5.5	···	2	*Pollard et al.* [1983]
	July 10–12, 1978	1	3	10	1	*Tryggvason* [1980]
	Sept. 4, 1984	0	2.5	9	1	*Tryggvason* [1986]
Etna, Italy	March 17, 1981	0	0.3	5	1	*Sanderson et al.* [1983]
	March 28, 1983	0–0.5	0.3–0.9	4	2	*Murray and Pullen* [1984]
	Sept. 28 to Oct. 2, 1989	0.2	1	7	1	*Bonaccorso and Davis* [1993]
	Dec. 14, 1991	0.45	0.6	3.5	1.3	*Murray* [1994]
Tolbachik, Kamchatka	July–Dec. 1976	0	···	10	1.1	*Fedotov et al.* [1980]
Kilauea, Hawaii	May 15–16, 1970	0.4	0.4	3	0.8	*Pollard et al.* [1983]
	Sept. 24–29, 1971	0	1.5	14	1.9	*Dvorak* [1990]
	Aug. 10, 1981	0.25	3	···	1	*Pollard et al.* [1983]
	Jan. 2, 1983	0	2.4	11	3.6	*Dvorak et al.* [1986]
	June 22–26, 1982	4	6	10	2	*Dvorak et al.* [1986]
Aira and Sakurajima, Japan	Jan. 12, 1914	0	2	7	8	*this paper*
Izu-Oshima, Japan	Nov. 21, 1986	0	2	4	1	*Hashimoto and Tada* [1990]
		2	10	12	2.7	
Teishi Knoll, Japan	July 1–11, 1989	1	6	3	1.2	*Okada and Yamamoto* [1991]
Miyakejima, Japan	1940	0	···	···	2	*Earthquake Research Institute* [1941]
Asal-Ghoubbet,	Nov. 6, 1978	0	4.5	4.5	2	*Tarantola et al.* [1979, 1980]
Djibouti		0	4.5	8	4	

intrusion the asymmetry of the two maxima indicates that the dike dips 85° west [*Pollard et al.*, 1983]. The amount of separation between the two maxima indicates the depth to the top of the dike. For the 1981 intrusion the dike rose to within a few hundred meters of the surface. In 1982 the dike rose to within about 4 km of the surface. A greater depth to the top of the 1982 dike is supported also by a deeper earthquake swarm that accompanied formation of that dike.

A single eruption of Etna illustrates the different displacement patterns produced by variable distance to the top of a dike [*Murray and Pullen*, 1984]. Earthquake locations suggest that the March 1983 eruption was fed by a radial dike that grew from a central magma core toward the south flank (Figure 7b). Two leveling routes crossed the dike, one near the summit at the 3000-m elevation and the other at the 2500-m elevation, the latter immediately upslope of the eruptive fissure. The profile of vertical displacement across the upper route indicates that the dike reached within 500 m of the surface (Figure 7c). The profile across the lower route indicates that the dike was within 100 m of the surface, as if the top of the radial dike was nearly horizontal and the eruptive fissure formed where a radial dike growing from a central column of magma within Etna intersected the south flank just below the leveling route (Figure 7d).

Three-Dimensional Sheet Models

At only a few volcanoes do geodetic networks extend beyond the surface limits of recently formed dikes, and so it is possible to determine the finite size of a dike in three dimensions. In these cases an appropriate model for a dike is a rectangular sheet. An analytic solution has been derived for the case of a sheet of any dip angle and constant opening across the entire sheet [*Okada*, 1985].

The dimensions of about two dozen dikes have been determined using the sheet model (Table 3). For the cases listed in Table 3 that have one or more parameters missing, a plane strain model was used to compute dike dimensions. This restriction was necessary because the network for those events did not extend far enough that all dimensions of the dike could be computed.

Each event in Table 3 is interpreted as the sudden emplacement of a single dike. An eruption occurred in those events in which the top of the dike is listed as 0 km. For most events a single dike thickness was determined. The exceptions are Izu-Oshima, Teishi Knoll, and Asal-Ghoubbet, in which the dike was subdivided into contiguous segments, each of which had a different width. The computed widths in Table 3 are within the range of 1- to 7-m dike widths revealed at eroded volcanoes [*Kuno*, 1964; *Gudmundsson*, 1983; *Walker*, 1987].

In a few cases it is necessary to model both the formation of a dike and the nearby subsidence over a shallow magma reservoir that supplied magma to the dike. The computed position of the top of a vertical sheet, representing the dike that supplied the east rift zone eruption of Kilauea in January 1983, coincided with the eruptive fissure if a point source model was also computed for subsidence of the summit area of Kilauea [*Dvorak et al.*, 1986]. Without the point source model the computed position of the sheet was 100 m north of the eruptive fissure. The simultaneous determination of a point source subsidence and a vertical sheet has been computed for several summit eruptions of Kilauea [*Yang et al.*, 1992].

When an eruption occurred, the computed dike length was often slightly longer than the eruptive fissure. For example, the 1983 eruptive fissure along the east rift zone of Kilauea was 8 km, and the computed dike length was 11 km. An extreme case was the November 1978 eruption along the Asal-Ghoubbet portion of the mid-ocean rift system. The length of the eruptive fissure was less than 1 km; however, the pattern of horizontal displacement showed that 2 to 4 m of opening had occurred over two segments of the rift system (Figure 8d). Presumably, magma was intruded beneath the entire segment, but it reached the surface along only a short part of the segment. Field observations of eroded dikes show that they erupt along only a part of their total length [*Casadevall and Dzurisin*, 1987].

Another application of the rectangular sheet model is the emplacement of a sill. *Ryan et al.* [1983] proposed that the elliptical shape of vertical displacement in the summit area of Kilauea might be caused by repeated filling, then withdrawal, of an elongated sill. For two periods of uplift in 1972 and 1973, they derived an aspect ratio of three for the horizontal dimensions of two separate sills, one located at a depth of 2 km and the other at 3 km. For both 6-month periods the amount of vertical opening was about 1 m.

A sill has also been proposed for the magma body that lies about 20 km beneath the Rio Grande rift near Socorro, New Mexico. If the body covers an area of 1200 km^2 [*Sanford et al.*, 1977] to 2000 km^2 [*Hartse and Sanford*, 1992] and if it is being supplied magma at the rate of 1–2 $km^3\ yr^{-1}$ [*Reilinger et al.*, 1980], then the sill has increased in thickness at a rate of 5–15 mm yr^{-1}. Leveling surveys suggest that the average uplift rate has been constant since 1912 [*Larsen et al.*, 1986]. If so, then the sill beneath Sorroco would have increased in thickness 0.4 to 1.2 m since the first leveling survey. Geomorphic evidence suggests that the average uplift rate has been constant for at least the last 20,000 years [*Ouchi*, 1983], which would correspond to an increase in thickness of 100–300 m.

WATCHING VOLCANOES GROW

The events described in this section illustrate some of the ways in which volcanoes grow and evolve via ground displacement. All possible ways have probably not yet been recognized because only a dozen or so volcanoes have been watched closely for a long period. Furthermore, the full range of possible volcanic activity revealed in the geologic record has not occurred at volcanoes where geodetic measurements have been conducted. For example, surface displacements have not yet been recorded during a major caldera-forming event, when ring dikes may form, or during the buildup to a major explosive eruption of a stratovolcano, such as the formation of Crater Lake, Oregon. Nevertheless, many patterns of surface displacement are common to the few volcanoes studied so far, and so we emphasize the common threads that link these volcanic systems.

Shallow Basaltic Reservoirs

The cycles of slow uplift followed by sudden subsidence recorded at the summit areas of the basaltic volcanoes Kilauea, Mauna Loa, and Krafla are evidence that magma repeatedly fills and then withdraws from a shallow reservoir beneath the summit of each volcano. The inferred cluster of point sources at each of these volcanoes locates a reservoir at a depth of about 3 km, the level of neutral buoyancy for a basaltic volcano [*Ryan*, 1987]. That depth corresponds to the top of ancient magma reservoirs revealed by deep erosion of intraplate volcanoes at Réunion [*Upton and Wadsworth*, 1969] and Saint Helena [*Daly*, 1927; *Baker*, 1968]. Such conformance suggests that the summit magma reservoirs in Hawaii and Iceland are active examples of the exposed intrusive bodies, which consist of layered basaltic intrusions surrounded by cross-cutting dikes and sills. The layered structure means that only part of the intrusion beneath a Hawaiian or Icelandic volcano is molten at any time. The surrounding dikes and sills explain the occasional horizontal migration of uplift and subsidence centers at these three volcanoes [*Fiske and Kinoshita*, 1969; *Tryggvason*, 1980] and the systematic filling and draining of the shallow reservoir beneath the summit area of Kilauea [*Waesche*, 1940].

Much of the elevation of the dissected volcanoes results from upheaval caused by intrusions. Though net upheaval is not evident in displacement measurements made at active basaltic volcanoes, vertical displacement of many tens of meters has been recorded within some explosive calderas, such as Usu, Japan [*Minakami et al.*, 1951; *Yokoyama et al.*, 1981]. In particular, during the 75 years that such measurements have been made at Kilauea, the summit region has had a net subsidence, mostly caused by a phreatic eruption in 1924 and the current eruption which began in 1983 (Figure 2d).

As the height of a volcano increases by continued eruption of lava flows, the magma reservoir grows upward, so that it remains at the level of neutral buoyancy [*Ryan*, 1987]. This is most strikingly revealed in Hawaii, where there is a 3-km difference in the summit elevations of Kilauea and Mauna Loa, but the depth below the summit of each magma reservoir is still about 3 km.

The summit areas of some large basaltic shield volcanoes subside slowly, sometimes without eruptions. Notable examples are Poas [*Rymer and Brown*, 1989], Colima [*Murray*, 1993], and Etna [*Murray et al.*, 1994]. Such slow subsidence might be caused by spreading of the volcano under gravity [*Borgia*, 1994] or loading of the crust by erupted lava [*Murray et al.*, 1994].

Further confirmation that magma accumulates at the level of neutral buoyancy is provided by exploration of the mid-ocean ridge system. Relative to the top of the oceanic crust, the level of neutral buoyancy along the mostly submarine segments of this volcanic system should be shallower than at on-land systems, such as Kilauea and Mauna Loa, because of the additional weight of the ocean over the submerged ridge system. Three kilometers of seawater is equivalent in weight to 1 km of rock, so the level of neutral buoyancy along a 3-km-deep segment of the ridge system should be about 1 km shallower than at Kilauea. This is confirmed by seismic imaging of the East Pacific Rise, where the top of a melt lens lies 1–2 km beneath the seafloor [*Detrick et al.*, 1987].

The lack of a large reservoir within a few kilometers of the summit of either Piton de la Fournaise [*Lenat et al.*, 1989b] or Etna [*Murray and Guest*, 1982; *Chester et al.*, 1985] is cited by those researchers as the reason for a fundamental difference in the behavior between these two basaltic volcanoes and the Hawaiian or Icelandic volcanoes. Several eruptions have occurred at both Piton de la Fournaise and Etna since intensive programs were started to measure surface displacements. The results revealed only small summit displacements, except in areas where displacements could be associated with emplacement of a radial dike.

On the other hand, the difficulty of measuring gradual summit uplift before recent eruptions of Piton de la Fournaise could be the result of a relatively low magma supply rate. If the rate is gauged by the volume of erupted lava, then since about 1950 the rate at Piton de la Fournaise has been only a tenth of the rate at Kilauea [*Lenat and Bachelery*, 1988]. Hence uplift rates might also be reduced by a tenth, even if the two volcanoes have summit reservoirs that are similar.

Furthermore, since 1979, when a geodetic program was started at Piton de la Fournaise, only one eruption occurred outside the summit area. When a summit eruption occurs at Kilauea, surface displacements are slight except around the eruptive fissure. The same has been true at Piton de la Fournaise. The sudden summit-wide subsidence recorded at Kilauea occurred during flank eruptions. During the only flank eruption monitored at Piton de la Fournaise, a 12-day eruption in March 1986 when 14 million cubic meters of lava was erupted, tilt changes were as large as 60 μrad [*Delorme et al.*, 1989]. The tilt pattern, however, indicated a pressure source within 1 km of the surface, significantly shallower than the 3-km-deep sources determined beneath the summits of Kilauea, Mauna Loa, and Krafla.

Summit earthquakes at these three volcanoes are within 3 km of the surface and cap magma reservoirs. At Piton de la Fournaise the maximum depth of summit earthquakes is also 3 km. By inference, the lack of deeper summit earthquakes could mean that a summit reservoir is present. Perhaps, when another major flank eruption occurs at Piton de la Fournaise comparable to those in 1931 or 1977, when the volume of erupted lava flows was 100×10^6 m^3 or more, the pattern of surface displacement might reveal a substantial magma reservoir located 3 km beneath the summit area of Piton de la Fournaise.

After more than 20 years of measurements at Etna, which spanned three voluminous eruptions of 79×10^6 m^3 in 1983, 62×10^6 m^3 in 1989, and more than 100×10^6 m^3 in 1991 [*Ferrucci et al.*, 1993; J. B. Murray, written communication, 1996], the patterns of surface displacement showed no evidence of a 3-km-deep reservoir. The axisymmetric patterns of slight uplift or subsidence around the summit area are evidence that only small magma volumes of the order of 0.1×10^6 m^3 are stored within a few kilometers of the surface [*Murray and Guest*, 1982]. Seismic tomography suggests a magma reservoir large enough to supply the volumes of recent eruptions at depths greater than 10 km [*Cardaci et al.*, 1993]. Earlier analyses of seismic signals suggested a large magma reservoir at a depth of 20 km [*Sharp et al.*, 1980]. If so, the removal of 100×10^6 m^3 from a 10-km-deep reservoir, based on a point source model, would produce a relative elevation change of only 20 mm between the summit of Etna and the reference bench mark located about 4 km from the summit. Such a small elevation change is difficult to measure using standard leveling techniques on the steep slopes of Etna. A significant difference between Etna and seafloor volcanoes is that Etna sits atop a 27-km-thick section of continental crust [*Sharp et al.*, 1980], which has a lower density than the oceanic crust that lies beneath Hawaii and Iceland.

The radial dikes that formed recently at Etna, most notably in April 1983 (Figure 7d), are features common at many eroded volcanoes, such as Sunlight volcano, Wyoming, and Mount Rainier, Washington. At Etna the radial dikes grew horizontally from a central magma core within the volcano. By inference, the central plugs at Sunlight and Rainier were also molten during growth of radial dikes at these volcanoes.

Rate of Magma Movement

The lack of a magma reservoir within or at the base of Etna means that the eruptive rate is a direct indicator of the magma flow rate through this volcano. From 1868 to 1950 the average rate was 0.005 km^3 yr^{-1}, after which it rose abruptly to 0.018 km^3 yr^{-1} [*Murray*, 1990]. A presumed 10-km-deep reservoir far beneath the volcano is beyond the sensitivity of the present geodetic networks, and so we do not know how the rate into or out of the reservoir may vary.

In Hawaii and Iceland, rates of magma movement into and out of shallow magma reservoirs can be determined from the amount and rate of surface displacement. After a sudden subsidence of the summit area of Kilauea or Krafla, the area rises at a relatively high rate, which decreases as the uplift progresses for a few weeks to a year or longer (Figure 10). Apparently, after sudden removal of magma from a reservoir, the rate of magma supply to the reservoir is fast initially and then decreases

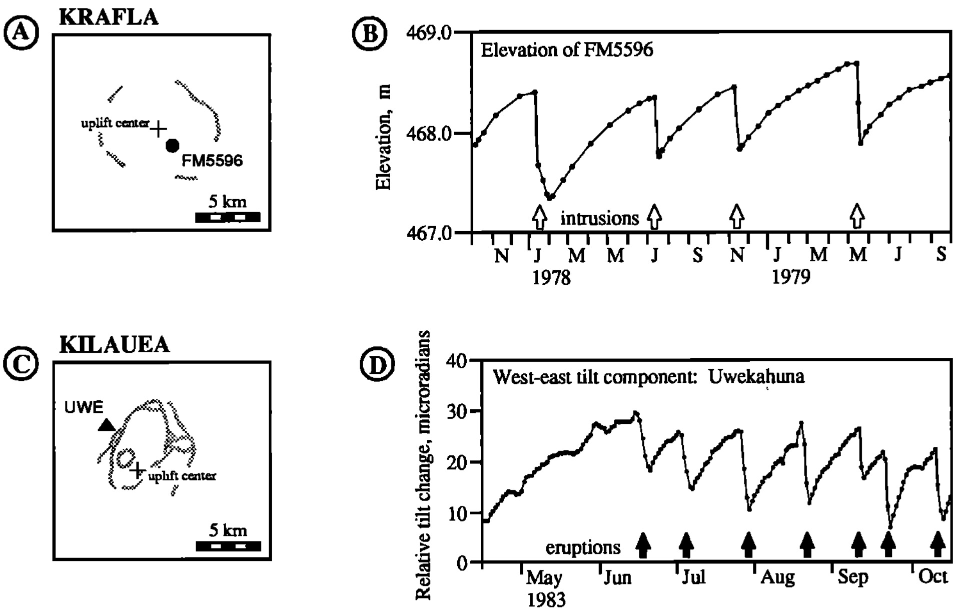


Figure 10. Change in uplift rate as magma enters shallow summit reservoirs at Krafla and Kilauea. (a) Location map of Krafla, showing an outline of the summit caldera as stippled lines and the location of the main uplift center and of bench mark FM5596. (b) Elevation change of FM5596 relative to a bench mark located 12 km to the south (modified from *Hauksson* [1983]). The timing of intrusions is indicated by vertical arrows. (c) Location map of Kilauea. (d) West-east tilt component of the water tube instrument in the Uwekahuna vault. The timing of eruptions along the east rift zone is indicated by solid arrows.

in a regular manner. The same change in displacement rate occurs during subsidence, when subsidence rate is high initially and then decreases (Figure 11). Similar variations in magma flow rate occurred during and after the 1984 eruption of Mauna Loa [*Lockwood et al.*, 1987].

The decrease of displacement rate can be approximated as an exponential decay, whether magma flows into or out of a reservoir. Such a response can be explained by assuming that the rate of magma flow is controlled by pressure difference between a summit reservoir and, during uplift, a magma source beneath the volcano or, during subsidence, supply of magma to a rift zone where it may be erupted [*Tryggvason*, 1986; *Dvorak and Okamura*, 1987]. A relationship between the amount of summit subsidence and the elevation difference between the site of the flank eruption and the summit area is further evidence that pressure of the summit reservoir controls magma movement. During a rift zone eruption of Kilauea, the volume of summit subsidence, and hence the volume of magma withdrawn from the summit reservoir, is generally larger when an eruption occurs at a lower elevation [*Epp et al.*, 1983]. A similar, though less well defined, relation between larger subsidence volumes and lower elevation of eruptions occurs at Krafla [*Björnsson et al.*, 1979; *Björnsson*, 1985].

Resurgent Domes

At many large silicic calderas, such as Yellowstone, Long Valley, Campi Flegrei, and Rabaul, geologic evidence indicates that a substantial amount of uplift has occurred near the center of each caldera since the caldera formed. The central parts of Yellowstone and Long Valley are occupied by resurgent domes (two at Yellowstone and one at Long Valley) that rise 100 m or more above the caldera floors [*Christiansen*, 1984; *Bailey et al.*, 1976]. At Campi Flegrei and Rabaul, oceanographic reflection surveys show that the caldera centers are arched upward in a similar way [*Pescatore et al.*, 1984; *Greene et al.*, 1986]. In addition, three marine layers, as old as 8400 years, are uplifted as much as 40 m at the center of Campi Flegrei [*Cinque et al.*, 1985].

The caldera centers are also the points where maximum uplift has been measured in recent years. The coincidence of past and present uplift near the caldera centers suggests a relationship between recent activity and the activity that formed the central uplifts, in particular, formation of resurgent domes. The relationship, however, is not direct. The profile of recent uplift is broader; it cannot be multiplied a hundred times or so to produce the topographic profile of the resurgent domes at Yellowstone (Figure 4) or Long Valley (Figure 12) or the local disturbances that are central to Rabaul or Campi Flegrei. How can this difference be explained?

Our explanation of recent uplift rests primarily on the abrupt start or end of some uplift episodes. In particular, episodes began within a month or less at Campi Flegrei in June 1982 [*Barberi et al.*, 1984], at Rabaul in 1983 [*McKee et al.*, 1985], and at Long Valley in October 1989 [*Langbein et al.*, 1993]. For Campi Flegrei and Rabaul, those episodes ended in less than a month, perhaps in as

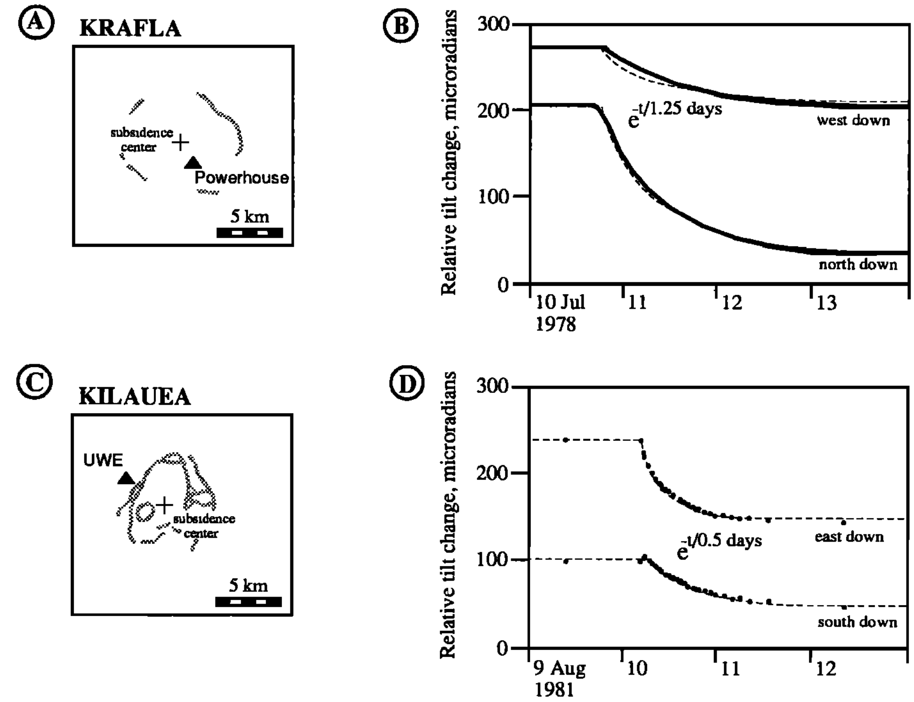


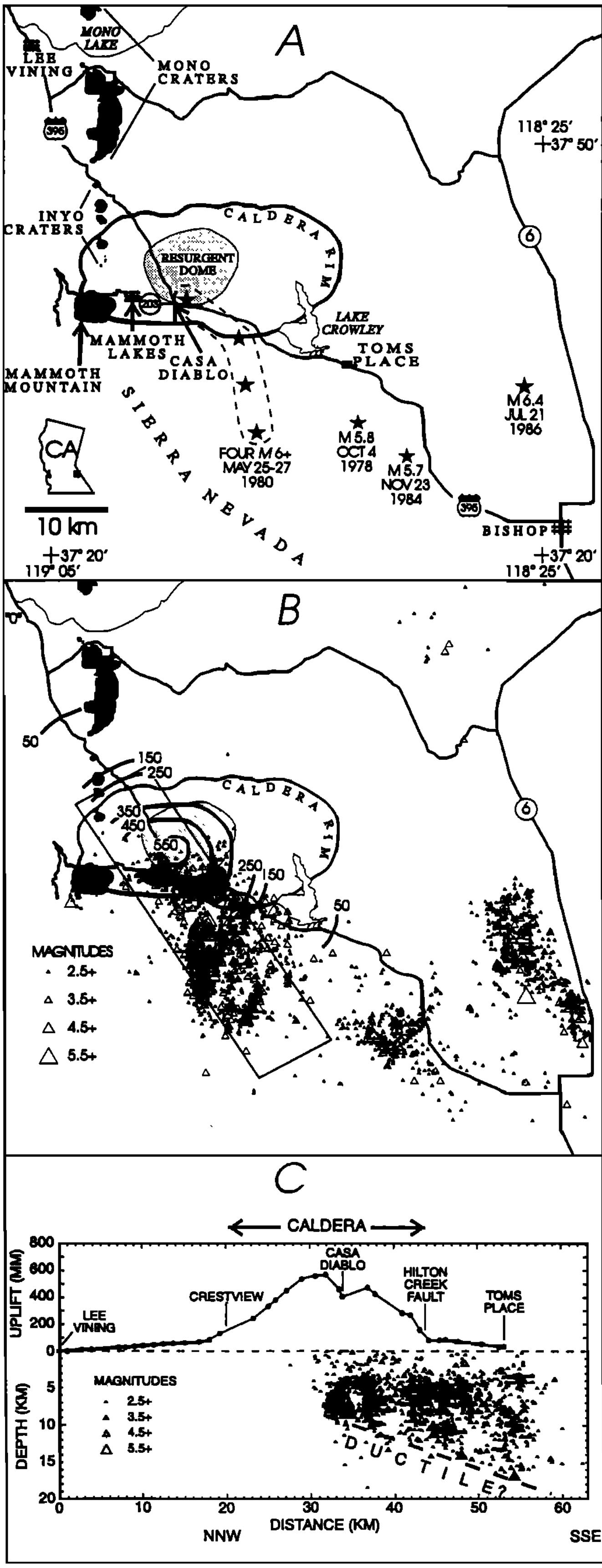
Figure 11. Change in subsidence rate as magma withdraws from shallow summit reservoirs at Krafla and Kilauea. (a) Location map of Krafla, showing outline of the summit caldera as stippled lines, the July 1978 subsidence center, and position of the Powerhouse electronic tiltmeter. (b) Tiltmeter record (solid line) and exponential decay curve (dashed line) for the July 10–12, 1978, intrusion into the north rift system [*Tryggvason*, 1980]. The slight difference between the tiltmeter record and the decay curve might be caused by a slow horizontal migration of the subsidence center. (c) Location map of Kilauea, showing the outline of the summit caldera as stippled lines, the August 1981 subsidence center, and position of the Uwekahuna vault, where a 3-m-long water tube tiltmeter is located. (d) Tilt readings (solid circles) and exponential decay curve (dashed line) for the August 10–11, 1981, intrusion into the southwest rift zone [*Dvorak and Okamura*, 1987].

little as a few days. For the following reasons, such rapid changes in displacement rate rule out the possibility that either internal pressurization of a hydrothermal system or a change in regional strain, for example by a large nearby earthquake, is the primary cause of uplift.

A hydrothermal system may be pressurized internally either by addition of magmatic fluids to a confined aquifer or by heating of fluids within a confined aquifer. The heat, ultimately derived from magma, is transferred by a combination of conduction and upward circulation of deeper and hotter fluids. Both the addition of fluids and the transfer of heat are diffusive processes, and so pressurization begins gradually, reaches a maximum rate, then decelerates. How fast are the transitions at the beginning and the end? The transitions are limited by the ability of fluids to change circulation patterns within the crust. The response of the levels of water wells to strain impulses suggests that these transitions occur over months or years [*Roeloffs et al.*, 1989; *Watanabe*, 1983], in contrast to the abrupt start or end of recent uplift episodes at Campi Flegrei, Rabaul, and Long Valley, which lasted less than a week.

Tectonic earthquakes have occurred months to years before the start of some, but not all, recent uplift episodes. Nineteen months before the beginning of uplift in June 1982, a $M = 6.9$ earthquake occurred 100 km south of Campi Flegrei; no comparable earthquake occurred before the uplift episode that began in about 1969. A $M = 5.7$ earthquake occurred 5 km south of Long Valley in October 1978. This was within 2 years of the first measurements that showed uplift of the caldera center. A $M = 7.6$ earthquake occurred 200 km east of Rabaul in March 1983, 6 months before uplift accelerated. These comparisons are suggestive but not conclusive, especially in the case of Campi Flegrei, where the first uplift episode was not preceded by a tectonic earthquake but the second episode was. Furthermore, no uplift episode at any of these three calderas began or ended within a month or so of the occurrence of a large tectonic earthquake, and so we doubt that such an earthquake was directly responsible for uplift.

Magmatic intrusion is the only known mechanism that could produce the abrupt start and end of the recent uplift episodes at these resurgent calderas. The ability of magma to move in a sudden manner is illustrated by the daily summit tilt record of Kilauea. Magma may intrude at a wide variety of rates, which include starting or stopping within a week or much less. Furthermore, we suggest that the proposed intrusions beneath the active



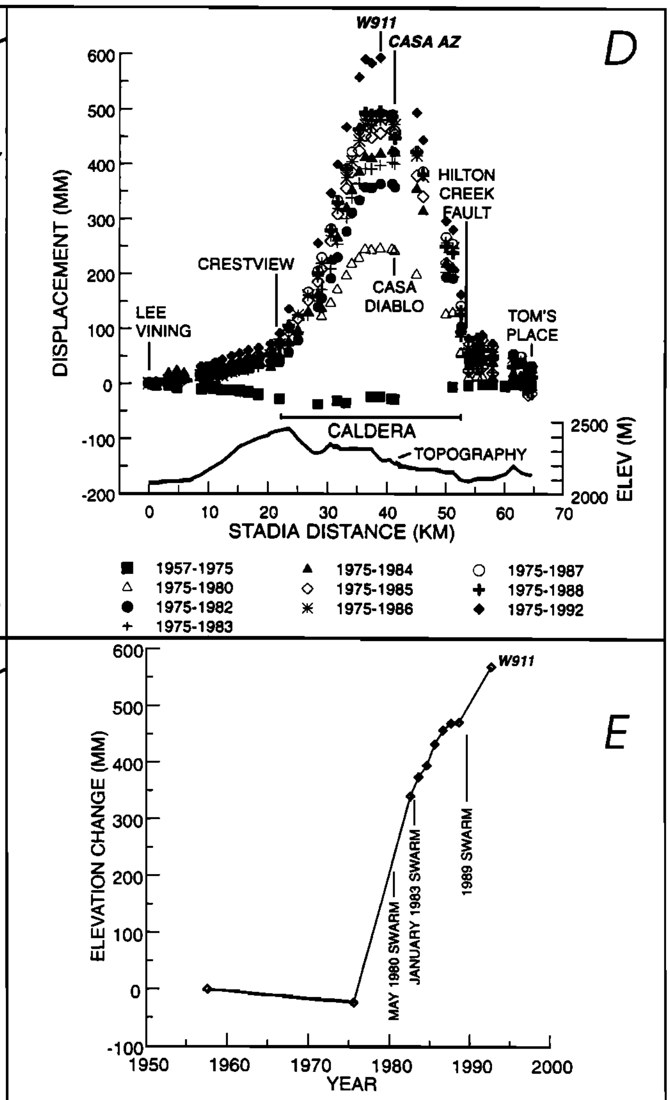


Figure 12. (a) Long Valley region, including Long Valley caldera, its resurgent dome, Mono Craters and Inyo Craters volcanic chains, and features mentioned in the text. Stars mark large earthquakes. (b) Locations of $M > 2.5$ earthquakes for the period 1980–1991 (triangles) and uplift contours (heavy lines, in millimeters) for 1975–1992. The rectangle encloses earthquakes shown in Figure 12c. (c) Uplift cross section along Highway 395 between Lee Vining and Toms Place for 1980–1992, plus hypocenters of earthquakes outlined in Figure 12b. Note the aseismic region beneath the caldera as in Figure 4c. Uplift is centered on the resurgent dome near Casa Diablo, with localized subsidence due to fluid withdrawal for geothermal power production. (d) Displacement between Lee Vining and Toms Place as measured by repeated leveling surveys. Bench mark W911, near Casa Diablo, is just outside the subsidence bowl; CASA AZ is in the subsiding area. (e) Elevation at bench mark W911 as measured by repeated leveling surveys. Uplift probably started about May 1980, when four $M > 6.0$ earthquakes occurred. The 1980 and 1983 earthquake swarms were beneath the south moat of the caldera, near Casa Diablo; the 1989 swarm was beneath Mammoth Mountain and was accompanied by increased uplift within the caldera.

resurgent calderas were basaltic magma injected into the bottom of silicic magma bodies.

In an earlier section of this paper, we recounted the view of many others that basaltic magma intrudes beneath silicic calderas. Specifically, intrusion of basaltic magma was invoked as the most efficient means of maintaining silicic magma beneath Yellowstone and Long Valley for periods of several hundred thousand years [*Lachenbruch and Sass*, 1978]. Also, basaltic lava flows have erupted peripheral to both Yellowstone and Long Valley [*Christiansen*, 1984; *Bailey et al.*, 1976]. Basaltic magma lies beneath Rabaul, as indicated by eruption of basalt since the most recent caldera-forming eruption 1400 years ago. The presence of basaltic magma beneath Campi Flegrei is problematic because volcanic activity since the formation of the caldera 32,000 years ago can be explained as the progressive crystallization of an isolated silicic magma system [*Armienti et al.*, 1983]. Trachybasalt, however, has erupted from the two volcanoes nearest Campi Flegrei, Ischia and Vesuvius, and so it is possible that basaltic magma rises also beneath Campi Flegrei.

The point source solutions to the uplift patterns at Campi Flegrei and Rabaul suggest basaltic magma rises to within a few kilometers of the surface (Table 2), a level that may be the point of neutral buoyancy, comparable to that beneath basaltic volcanoes. Beneath the larger calderas of Yellowstone and Long Valley, if intrusion of basaltic magma is the cause of uplift, then that magma stalls at depths deeper than about 5 km (Table 2), slightly shallower than the estimated depth that basaltic magma rose into the Cordillera del Paine granite [*Michael*, 1984]. Basaltic magma stalls because it encounters a large silicic magma chamber of lower density [*Eichelberger and Gooley*, 1977] or, perhaps a transition occurs from transport via magma fracture to basaltic fluid flow through a viscous silicic magma [*Valentine*, 1993].

Because the profile of a resurgent dome is narrower than the recent uplift profile at either Yellowstone or Long Valley, a resurgent dome forms by a process that acts at a shallower level in the crust than our proposed intrusions. Dome growth may be episodic, as indicated by marine layers exposed in a sea cliff in the center of Campi Flegrei. The uppermost and youngest marine layer was uplifted at least 40 m during a 2500-year period of dormancy [*Cinque et al.*, 1985].

A resurgent dome may grow by slow arching of the top of a large silicic magma chamber, evident at eroded explosive calderas such as in the San Juan volcanic field [*Steven and Lipman*, 1976]. The arching may be caused by upward movement of a silicic magma body, which could be episodic if it is heated by occasional deeper basaltic intrusions that are fed magma at variable rates. An example of such episodic movement is a resurgent dome within the explosive caldera at Iwo Jima, where shoreline terraces and results of leveling surveys show uplift rates have varied from 100 mm yr^{-1} to almost 500 mm yr^{-1} during at least the last 700 years [*Kaizuka et al.*, 1983, 1985].

Long-Term Subsidence of Calderas

Some calderas that formed over shallow basaltic reservoirs or over silicic magma chambers subside at relatively slow rates for long periods of time. Examples are Campi Flegrei [*Berrino et al.*, 1984], Askja [*Camitz et al.*, 1995], Aira [*Tada and Hashimoto*, 1989], Yellowstone [*Dzurisin et al.*, 1990], and Medicine Lake [*Dzurisin et al.*, 1991]. Only at Aira has long-term subsidence occurred during lateral magma movement from beneath the caldera to feed intermittent eruptions of Sakurajima. The other examples have no clear evidence of concurrent magma withdrawal from beneath the caldera. Such evidence might have been shallow earthquake swarms or simultaneous uplift over a nearby area of the presumed magmatic intrusion.

Subsidence may persist for a few years or may last a thousand years or longer, interrupted by uplift that may presage an eruption. Patterns of vertical displacement show that maximum subsidence occurs near the caldera center at rates of 10 to 40 mm yr^{-1}. At the two examples where uplift has also occurred recently, Yellowstone (Figure 4) and Campi Flegrei [*Dvorak and Berrino*, 1991], the profiles of uplift and subsidence are mirror images. This suggests that the source of subsidence is located at or very near the source of uplift.

If magma bodies have persisted beneath each of these calderas at least since formation of the caldera, then the subsidence rates are too high to be caused solely by cooling and crystallization of magma, for example at Campi Flegrei [*Dvorak and Mastrolorenzo*, 1991].

Most calderas that subside slowly lie within areas of regional extension, and so some subsidence might be caused by thermal weakening of the crust by a magma body beneath the caldera, for example at Yellowstone [*Dzurisin et al.*, 1990] and Medicine Lake [*Dzurisin et al.*, 1991]. The alternative, however, can explain neither the accelerated subsidence rate at Campi Flegrei after both recent episodes of rapid uplift nor the onset of subsidence at Yellowstone during 1985–1986.

While local contraction by cooling and regional extension may contribute to subsidence, the dominant factor is probably slow removal of magmatic fluids as the magma body cools and releases fluids to the overlying hydrothermal system. Both uplift episodes since 1968 at Campi Flegrei were followed by subsidence at a rate that was larger than the average rate during the previous 140 years [*Berrino et al.*, 1984; *Barberi et al.*, 1984]. The increased subsidence rate slowly decayed, so that after a few years it had returned to the long-term rate. A similar response was recorded by leveling surveys across a presumed magmatic intrusion in 1966 during an earthquake swarm beneath Matsushiro on the Izu Peninsula, Japan. (An alternative explanation is that the event was the result of dilation related to slip along a horizontal fault [*Nur*, 1974].) The rate of surface water flow increased during and after the event; the volume of water outflow after the end of the intrusion was about equal to the volume of subsidence [*Kisslinger*, 1975].

Growth of Eruptive Fissures

Most patterns of surface displacement presented so far were obtained by resurveys of extensive geodetic networks. Because each survey required at least several days, and often several weeks, to complete, the results yield only a net change in displacement after an eruption. No information can be gleaned about how the eruption developed and progressed, in particular the rate at which an eruptive fissure formed. This has changed in the last 20 years, so that by using automatic instruments, such as tiltmeters and strain meters, it has been possible to record formation of eruptive fissures, and surface displacements during dangerous explosive eruptions. The latter will be considered in the next subsection.

The growth of the January 1983 eruptive fissure of Kilauea was recorded by six tiltmeters located along the axis of the east rift zone. During a 24-hour period the epicenters of shallow earthquakes expanded along the axis of the east rift zone at a rate of 0.6 km hr^{-1} [*Koyanagi et al.*, 1988]. Also, a migration was noted of the timing of initial surface displacement recorded by the tiltmeter network [*Okamura et al.*, 1988]. The tilt records can be used to determine the amount and rate of rifting that occurred within the volcano before the eruption. As an example, consider the complete record of the Kamoamoa tiltmeter.

The Kamoamoa tiltmeter was located about 300 m north of the growing east-west dike that would feed the eruption (Figure 13a). The tilt record is shown as a component perpendicular to the rift axis and another component parallel to the rift axis (Figure 13b). The beginning of the shallow earthquake swarm and the onset of the eruption 24 hours later are indicated as vertical dashed lines.

The growing dike is modeled as a vertical sheet that begins at a point at a depth of 2 km, then grows horizontally at a constant rate of 0.5 km hr^{-1}, about the same as the earthquake migration rate, and vertically at a constant rate of 0.08 km hr^{-1} (Figure 13b). The width of the sheet is always 2.5 m. In this model the sheet continues to grow until it intersects the surface at the onset time of the eruption. The model reproduces the basic features of the tilt record, such as the delayed and slow turnover of both tilt components, the amount of tilt change of both components, and the transient change of the parallel tilt component. The large difference between computed and recorded tilt for the perpendicular component, which begins about 3 hours before the eruption, may indicate inelastic displacement or a rapid change in dip as the top of the dike nears the surface.

The complexity of the Kamoamoa tilt record can be understood better by viewing the evolving pattern of cumulative elevation change that occurred as the sheet continued to grow. Again, the sheet is assumed to begin at a point at a depth of 2 km (Figure 14a). After 3 hours, two ridges are apparent in the pattern, separated by a trough above the growing sheet (Figure 14b). During the next 15 hours the two ridges grow in amplitude, elongate parallel to the rift axis, and move closer to the rift axis (Figures 14c, 14d, and 14e). Finally, as the sheet touches the surface, the ridges merge, and the eruptive fissure forms along the crest of a single long ridge, which, in this example, was uplifted a maximum of 0.5 m (Figure 14f).

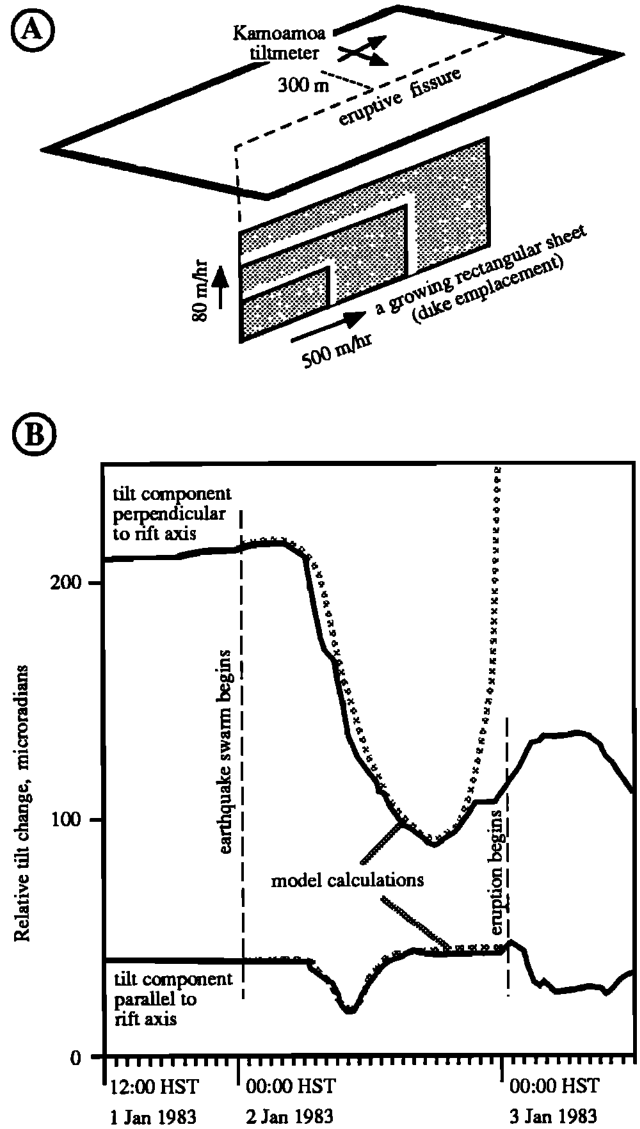


Figure 13. Growing pressurized dike, used to model the tilt changes that occurred as an 8-km-long eruptive fissure developed along the east rift zone of Kilauea on January 2, 1983. (a) Diagram showing location of the tiltmeter about 300 m from the eventual eruptive fissure. The model assumes that the dike grows at constant rates vertically and horizontally and takes the shape of a vertical, rectangular sheet. (b) Comparison of measured (solid lines) and computed (dashed lines) tilt changes for the growing dike. (Modified from *Okamura et al.* [1988].)

Besides Kilauea, the horizontal growth rate of dikes

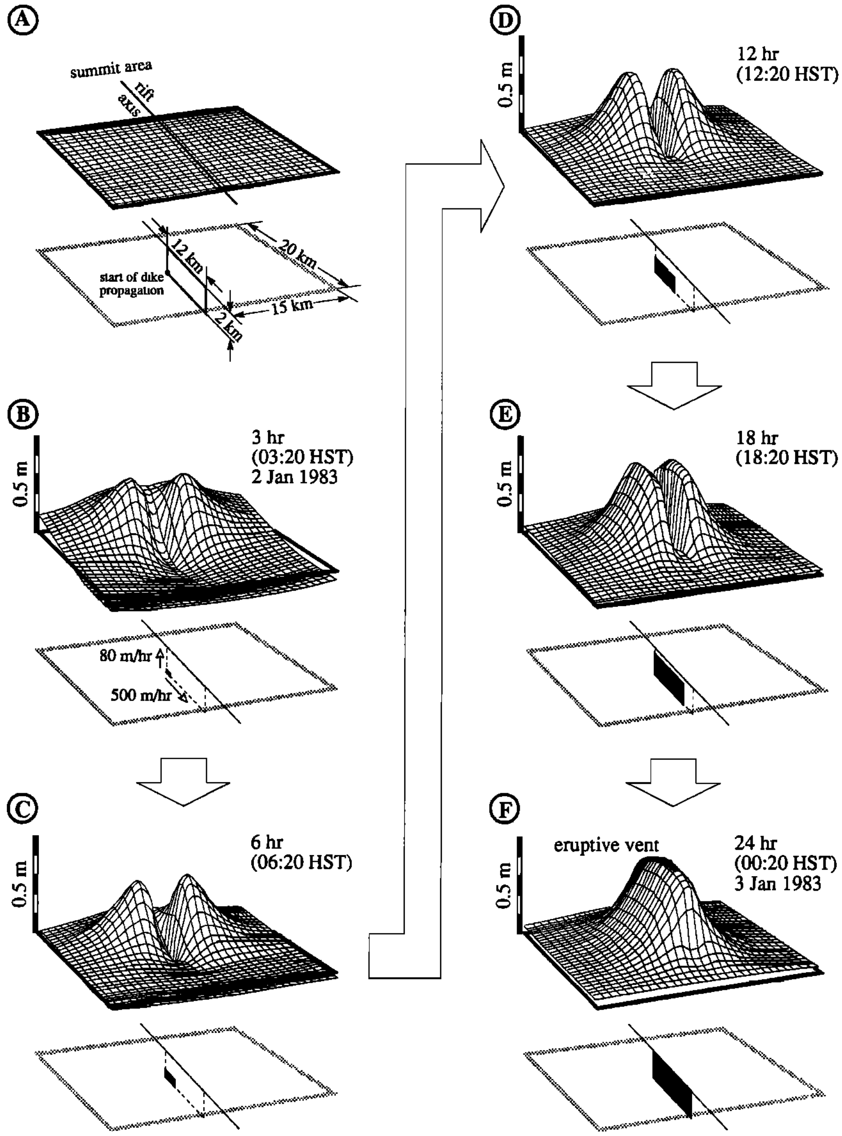


Figure 14. Elevation changes during the growth of a dike and eruptive fissure along the east rift zone of Kilauea on January 2, 1983. (a) Onset of dike growth. The dike is assumed to begin at a point, then grow at a constant rate of 80 m hr^{-1} vertically and a constant rate of 500 m hr^{-1} horizontally. (b) Three hours after the onset of dike growth. Two maxima form on either side of the axis of the rift zone, and a small trough forms along the axis, over the growing dike. As the dike nears the surface, these maxima will move closer until they finally merge when the dike reaches the surface (f). (c) Six hours after onset. (d) Twelve hours after onset. (e) Eighteen hours after onset. (f) Twenty-four hours after the onset of dike growth. The dike has reached the surface, and an eruptive vent has formed across the crest of the uplifted surface.

has been measured at a few other basaltic volcanoes and, recently, along a section of the Juan de Fuca ocean ridge (Table 4). All such determinations, whether from tilt records, earthquake migration rates, or actual observation of the rate at which surface cracks open, yield about the same rate of several hundred meters per hour up to 2 km hr^{-1}. These rates are much slower than the few kilometers per second associated with seismic velocities or the growth rate of cracks in ceramics and other brittle material. Therefore the rate of dike growth is not controlled by the rate at which a crack can form, but by the ability of viscous magma to keep up with a growing crack [*Spence and Turcotte*, 1985].

Explosive Eruptions

Measurements of surface displacement during explosive eruptions is among the most difficult tasks to perform around active volcanoes. Obviously, in view of the fatal disaster at Galeras, Colombia, in January 1993, one difficulty is the danger of people being too close and thus at high risk. Even when automatic instruments are used, a remaining problem is to locate an instrument close enough to record surface displacements but not so close that it will not survive an eruption. For these reasons, records of explosive eruptions have been made at only a few volcanoes, but these records reveal important characteristics about the physical nature of those explosions.

As an example, the tilt records and the histories of tremor amplitude are shown in Figure 15 for two explosive eruptions of Mount St. Helens in 1980. For each example the timing of the major explosion is indicated as a vertical line. The tilt records are shown as a component radial to the volcano, such that an increase in radial tilt represents uplift of the volcano, and a component tangential to the volcano.

In both cases, when tremor amplitude increased suddenly, the volcano subsided for an hour or two, then recovered to the original level during the next several hours. Apparently, magma from a shallow source was erupted suddenly, causing the initial subsidence, then

TABLE 4. Growth Rate of Dikes

Volcano	*Date of Event*	*Method*	*Horizontal Distance, km*	*Rate, km h⁻¹*		*Reference*
				Horizontal	*Vertical*	
Piton de la Fournaise, Réunion	April 18, 1990	changing tilt directions	4	0.7	···	*Toutain et al.* [1992]
Kilauea, Hawaii	Sept. 21–24, 1971	opening ground fissures	12	0.6–4.0	···	*Duffield et al.* [1982]
	Jan. 2, 1983	earthquake migration	10	0.6	···	*Koyanagi et al.* [1988]
		tiltmeter curve		0.7	0.07	*Okamura et al.* [1988]
Mauna Loa, Hawaii	March 25, 1984	opening ground fissures	20	1.2	···	*Lockwood et al.* [1987]
Krafla, Iceland	Sept. 8, 1977	earthquake migration	10	1.8	···	*Brandsdóttir and Einarsson* [1979]
		opening ground fissures		1.8	···	*Björnsson et al.* [1979]
	July 10, 1978	earthquake migration	30	1.6	···	*Einarsson and Brandsdóttir* [1980]
	Sept. 8, 1977	extensometer network	10	2.1	···	*Hauksson* [1983]
	March 16, 1980	extensometer network	7	1.7	···	*Hauksson* [1983]
Miyakejima, Japan	Oct. 3, 1983	opening ground fissures	3	1.5	···	*Aramaki et al.* [1986]
Teishi Knoll, Japan	July 13, 1989	earthquake migration	···	···	0.01–0.4	*Ueki et al.* [1993]
Juan de Fuca Ridge	June 26, 1993	earthquake migration	20	1		C. Fox (oral communication, 1994)

was replaced within several hours by magma from a deeper source. No net surface displacement resulted after an explosive eruption. A similar result of no net displacement has been recorded at two other exploding volcanoes.

The tilt and strain records during small explosions of Sakurajima show a transient change but no net change (Figure 16c). Surface strain changes occurred no more than a fraction of an hour before an explosion, offering no reasonable time to forewarn people around the volcano. A similarly short precursory time before explosions was also observed at Galeras, Colombia.

A tiltmeter was located within a kilometer of the explosion crater at Galeras (Figure 17b). For three-and-a-half months in 1989, the tiltmeter recorded two dozen spikes (Figure 17c), each with a rapid beginning and slower recovery, and no net change in ground tilt (Figure 17d). Each event is thought to be caused by sudden gas emission. As at Sakurajima, precursory strain occurred a fraction of an hour before an eruption, and no net change was observed a few hours after the explosion had occurred.

At Sakurajima and Galeras the tilt records might be the result of the sudden release of gas, which expanded rapidly as it rose the last several hundred meters to the surface and then was released as the ground surface returned to its preexplosion position. No significant volume of magma was involved. For Mount St. Helens, magma was erupted, so the transient change in ground displacement was not caused solely by gas release. Magma from a deeper level must have moved up rapidly to replace magma just erupted from a shallow reservoir. The rapid replacement of magma has made measurements of net surface displacements of limited use in predicting explosive activity at stratovolcanoes. Besides the three stratovolcanoes already mentioned, a similar result of no measured surface displacement after an explosive eruption has been noted during the 1986 eruptions of Nevado del Ruiz [*Banks et al.*, 1990] and the 1992 eruption of Manam [*McKee et al.*, 1993]. Adequate tracking of ground displacements at volcanoes such as these requires continuous measurements by tiltmeters, strain meters, or GPS instruments.

Predicting Volcanic Activity

Though measurements of surface displacement before explosive eruptions have been difficult to obtain, such measurements, along with earthquake patterns, have been very successful indicators of other types of eruptions. One of the best records of successful predictions has been at Mount St. Helens after the cataclysmic eruption of May 18, 1980 (Figure 18) [*Swanson et al.*, 1983].

The successful predictions had three key ingredients: (1) a volcano that had a series of eruptions, which over a few years changed in a systematic way from dramatic explosions to quiet dome extrusions, (2) ability to process seismic signals immediately, and (3) intensive fieldwork, usually done daily or weekly, during which simple, repeated measurements were made of surface displacement. The predictions of explosive eruptions relied on seismic data alone. Later, predictions of episodes of dome growth took advantage of the development of structural features around the growing dome, in particular the development of radial cracks and concentric thrust faults near the base of the dome.

Movement of thrust faults around the base of the dome accelerated a few days to weeks before an extrusive episode [*Chadwick et al.*, 1983]. Tilting of the crater floor also accelerated before several episodes [*Dzurisin et al.*, 1983]. Before each episode a prediction window was announced to the public. As measurements contin-

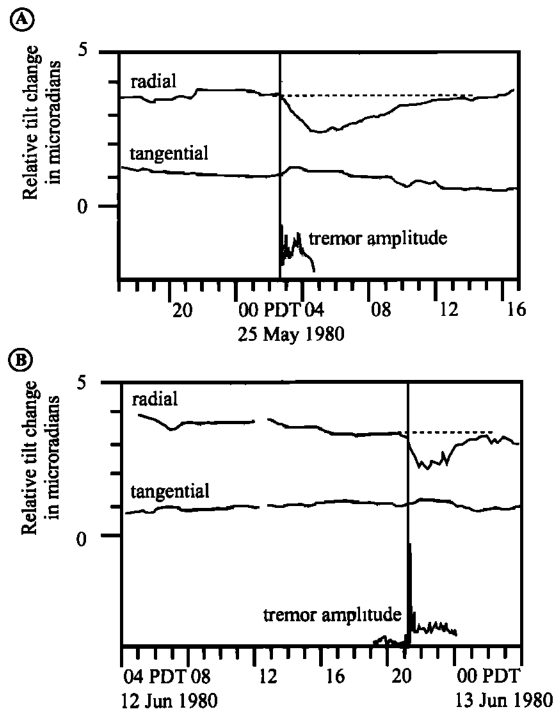


Figure 15. Ground movement and tremor amplitude during explosive eruptions of Mount St. Helens. Records of ground movement are from an electronic tiltmeter 7 km south-southwest of the eruptive vent [*Dvorak et al.*, 1981]. Tilt components are either radial or tangential to the vent. Records of tremor amplitude are from a seismic station located 16 km from the eruptive vent [*Scandone and Malone*, 1985]. The thin vertical line indicates the beginning of a major explosive column. (a) May 25, 1980. (b) May 12, 1980. Note that little, if any, net tilt change occured several hours after each explosive eruption.

ued to be made, the window was revised and narrowed. In this manner, seven successive dome-building eruptions were predicted between 3 days and 3 weeks in advance (Figure 18) [*Swanson et al.*, 1983].

Another example of how a repeated pattern of surface displacement is related to episodic eruptions is the four dozen eruptions that occurred along the east rift zone of Kilauea from 1983 to 1986. Most episodes, which lasted about 1 day each, were preceded by a few weeks of decreasing uplift rate in the summit area, indicated by a continuously recording tiltmeter [*Wolfe et al.*, 1987], and an increase in the rate of shallow summit earthquakes. Often, the start or end of an eruption along the east rift zone occurred within a few hours of the start or stop of summit subsidence. Immediately after an eruption, uplift of the summit area resumed; then over the next few weeks, an increase in shallow earthquakes occurred, and eventually another eruption began. The simple, repetitive curves of summit tilt during subsidence at both Kilauea and Krafla (Figure 11) suggest that the ability to predict the maximum duration and magnitude of a subsidence event at these volcanoes is possible.

The familiar sequence of initial ground movement followed by the start of shallow earthquakes has also been observed at other volcanoes. For example, uplift at Campi Flegrei that began in about June 1982, as measured by a tide gage, preceded by about 9 months the onset of a series of damaging earthquake swarms be-

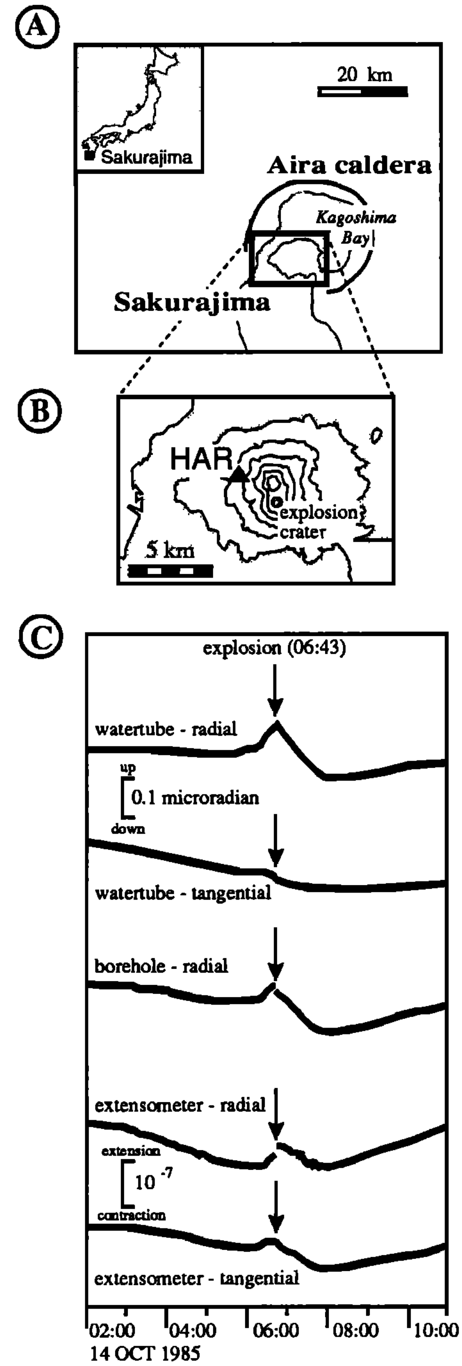


Figure 16. Sudden ground movement at Sakurajima, Japan. (a) Location map of Sakurajima, on the southern rim of Aira caldera. (b) Location of continuous recording instruments (HAR), about 2 km northwest of the explosion crater. (c) Continuous records of ground movement associated with a small explosion at 0643 LT on October 14, 1985. (Modified from *Ishihara* [1990]. Copyright John Wiley and Sons Ltd. Reproduced with permission.) A small uplift and radial extension of the volcano occurred 1–2 hours before the explosion.

neath the caldera center (Figure 19a) [*Barberi et al.*, 1984]. Both the uplift and earthquake swarms stopped in December 1984. A recent increase in surface strain rate at Long Valley was followed 2 months later by an increase in rate of shallow earthquakes (Figure 19b) [*Langbein et al.*, 1993]. Such a sequence is easy to understand. First, the crust starts to expand as magma intrudes beneath the volcano; then when the amount of strain exceeds a threshold value, of the order of tens to hundreds of parts per million, the crust starts to fracture, and earthquakes occur. If the process continues, a fracture eventually forms between the magma body and the surface, and an eruption ensues.

The relation between the onset of ground displacement and seismicity, however, is at other times unclear. The uncertainty may represent a lack of accurate and frequent measurements of surface displacements rather than a different volcanic pattern. As an example, consider the tantalizing result recorded during the last 30 years when two eruptions occurred at Mauna Loa in Hawaii (Figure 19c) [*Lockwood et al.*, 1987]. After 10 years of no significant ground displacement, a series of shallow earthquake swarms began in December 1974, 7 months before the July 1975 eruption. Extension across the caldera began at least by February 1975 and perhaps as early as November 1974. Before the next eruption in March 1984, the rate of shallow earthquakes increased gradually during the few years before the eruption, then dramatically an hour before the eruption. Extension across the caldera was also gradual a few years before the 1984 eruption. A tiltmeter in the summit area showed that rapid changes began an hour before the eruption, when the rate of shallow earthquakes also increased. This is an example in which the dual techniques of seismic and geodetic studies yielded a quantitative way to follow the buildup and predict the 1984 eruption [*Decker et al.*, 1983].

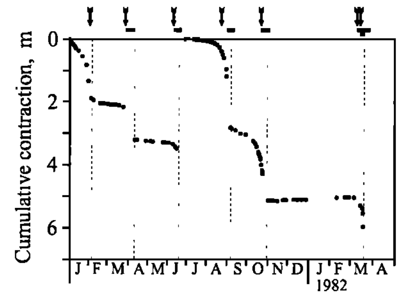


Figure 18. Increasing surface strain rate before periods of dome growth at Mount St. Helens (modified from *Swanson et al.* [1983]). Shown here are changes in distance across thrust faults near the dome. Vertical dashed lines indicate the timing of episodic dome growth. The bars at the top show the windows of predictions issued publicly before eruptive activity. The arrows show when the prediction was issued. Two predictions were issued in March 1982; the narrower predicted window was an update issued a few days before activity.

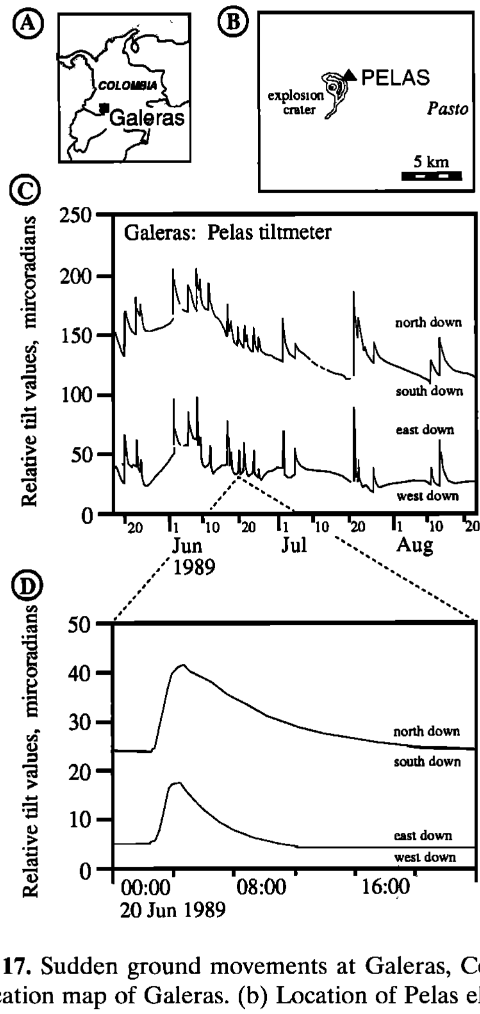


Figure 17. Sudden ground movements at Galeras, Colombia. (a) Location map of Galeras. (b) Location of Pelas electronic tiltmeter, installed within 1 km of the explosion crater on May 10, 1989. (c) Four-month record of tilt changes recorded by the Pelas tiltmeter. North or east down corresponds to uplift of the volcano. The event on May 20 was coincident with sudden emission of sulfurous gas. Other transient events are possibly related to gas emission. (d) Twenty-four-hour record of one transient tilt event, presumably related to sudden gas emission from the explosive crater [*Lockhart et al.*, 1990].

THE COMING REVOLUTION

The buildup of volcanic deposits by eruption of lava does not completely explain the growth of all volcanoes. In some cases, such as recent activity at the explosive calderas of Iwo-Jima and Usu and along the portion of the mid-ocean ridge at Asal-Ghoubbet, volcanic systems are growing faster by intrusion than by extrusion. Though basaltic volcanoes, such as those in Hawaii and Iceland, are composed of thousands of thin lava flows, the structure of a basaltic volcano is controlled not by the distribution of lava flows, but by magma accumulation in shallow reservoirs and by repeated dike intrusions. The cumulative effect of many dike intrusions alters the strain distribution within the volcano, which

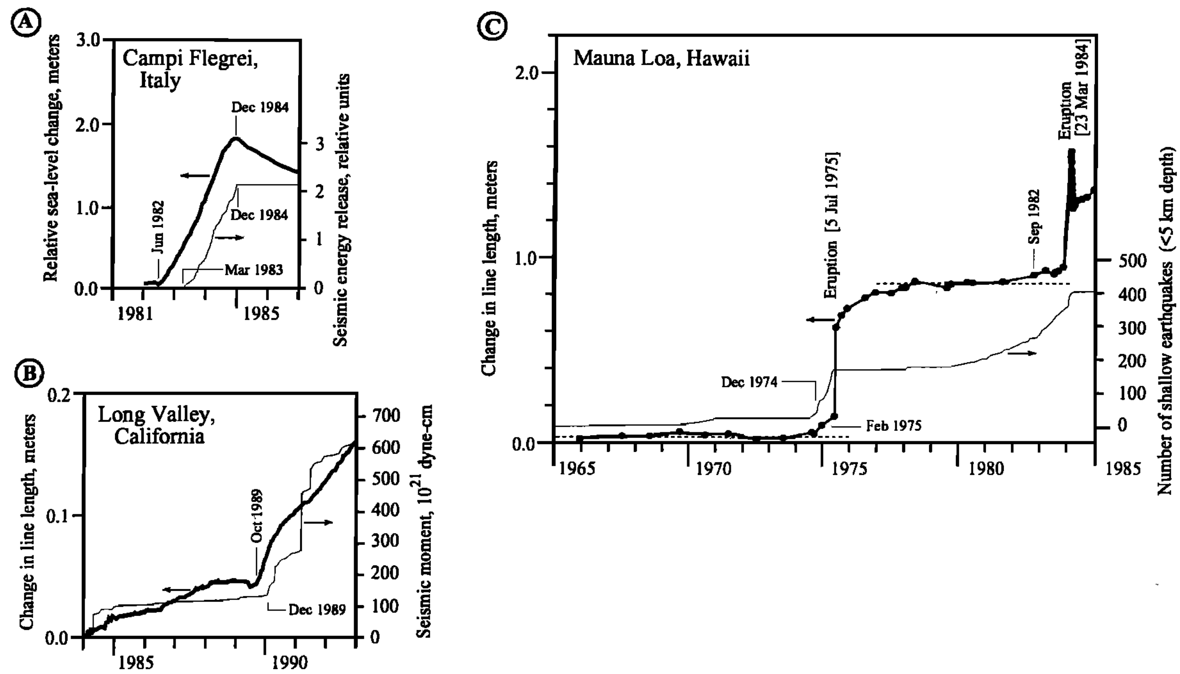


Figure 19. Time relation between start of ground movement and shallow seismicity. (a) Campi Flegrei. A tide gage record (thick line) near the point of maximum uplift shows that uplift began in June 1982 and ended in December 1984, followed by subsidence at a slower rate. Seismic energy release and earthquake rate began in March 1983 and ended in December 1984. (Modified from *Barberi et al.*, [1984].) (b) Long Valley. Daily measurements by a two-color laser instrument of changes in line length across the resurgent dome indicate that the rate of uplift increased in October 1989 and remained nearly constant for at least 3 years. The cumulative seismic moment and earthquake rate increased in December 1989. (Modified from *Langbein et al.* [1993].) (c) Mauna Loa. Yearly trilateration measurements across the summit caldera recorded about 200 mm of extension before the last two eruptions, in 1975 and 1984. The sudden extension during an eruption is caused by the spreading open of the surface as an eruptive fissure and dike formed. The sudden contraction after the start of the 1984 eruption records summit-wide subsidence. Dashed horizontal lines are average values of the line length during periods of no activity before each eruption. Significant changes in line length were measured in February 1975, about 7 months before the 1975 eruption and in September 1982, about 2 years before the 1984 eruption. The earthquake rate increased noticeably in April 1974 and dramatically in December 1974. The 1984 eruption was preceded by 4 years of gradual increase in earthquake rate; a dramatic increase did not occur until a few hours before the eruption. The thin line shows the cumulative number of shallow earthquakes that could be located. (Modified from *Lockwood et al.* [1987].)

changes the likelihood that a dike will reach the surface and feed an eruption, and eventually may lead to a large landslide, as at Kilauea [*Decker*, 1987]. These internal contributions to growth of an active volcano are best revealed by the patterns of surface displacement.

Besides surface displacement that is local to a volcano, displacement may also occur on a regional scale, a result of activity of a broad magmatic system that feeds several individual volcanoes. For example, the upward growth of a batholith may raise an entire volcanic field, such as the San Juan field [*Lipman*, 1984]. Also, permanent uplift may result by accumulation of a broad basaltic intrusion at the base of the crust (a process called underplating) such as one proposed along a segment of the trace of the Hawaiian hot spot [*Watts et al.*, 1985] or another beneath the Basin and Range province of the western United States [*Jarchow et al.*, 1993]. Underplating may have raised the Colorado Plateau [*McKenzie*, 1984] or eastern Australia [*Wellman*, 1989], both areas of ancient volcanism. The passage of a tectonic plate over a hot spot may raise and then lower a broad region. For example, the passage of the North American plate over the Yellowstone hot spot was invoked to explain a reversal of regional drainage patterns in Montana [*Fritz and Sears*, 1993]. The various components of regional displacement that are suggested by geologic evidence have not been measured yet using geodetic techniques, with one notable exception. The 40-year record of a tide gage on the island of active volcanoes in Hawaii has indicated continuous and slow subsidence. The subsidence may be an isostatic adjustment of the crust as the volcanoes continue to erupt and grow [*Moore*, 1987]. An alternative explanation is that the tide gage is located at a point that is drifting away from the maximum uplift of the crust caused by the Hawaiian hot spot. In essence, according to the second explanation, movement of the

Pacific plate causes the tide gage to "slide down" the hot spot swell.

These examples illustrate the need to expand the current coverage of geodetic measurements to many other volcanoes and show that the action of individual volcanoes cannot be divorced from regional activity. Just as astronomers succeeded in understanding the evolution of stars by the study of thousands of stars, instead of the few dozen that are nearest to us, earth scientists will better understand the evolution of magmatic systems when hundreds of active volcanoes are studied intensely, instead of the few dozen studied now. Such a major advance requires the use of new techniques that are able to make more frequent and more exact measurements of surface displacement than are possible with conventional techniques.

One new technique suited especially well to the study of active volcanoes relies on the satellite-based Global Positioning System. The advantages of GPS over conventional high-precision geodetic techniques is that GPS can be used in remote and rugged terrain and can record displacements of a few millimeters across baselines of a few kilometers to hundreds of kilometers [*Dixon*, 1991]. Already GPS surveys recorded displacements at Izu-Oshima [*Okada and Yamamoto*, 1991], Etna [*Nunnari and Puglisi*, 1994], and Kilauea [*Dvorak et al.*, 1995]. Permanent GPS stations that send data to a central collecting site offer another way to make continuous measurements of displacement, which are important for recording the development of a volcanic vent immediately before an eruption. Already such stations are used to record daily tectonic strain in southern California [*Bock*, 1993] and British Columbia [*Dragert et al.*, 1993] and continuous uplift and horizontal displacement at Long Valley [*Webb et al.*, 1995].

Another technique for making continuous measurements is the use of borehole strain meters, which are able to record strains as small as one part in 10^{-12}. A network of such strain meters recorded the simultaneous growth of an eruptive fissure and withdrawal of magma from a shallow reservoir beneath Hekla [*Linde et al.*, 1993]. Because borehole strain meters have an ultrahigh sensitivity, the wider use of these instruments could reveal events at distant volcanoes, such as the event near Matsushiro, Japan, in 1966 [*Stuart and Johnston*, 1975] and beneath Efate Island, New Hebrides, in 1986 [*Mellors et al.*, 1991], which were interpreted as magmatic intrusions.

Radar interferometry, either from an airplane or a satellite, offers the possibility to measure the two-dimensional pattern of vertical displacement caused by magmatic or seismic activity, instead of the scattered point measurements made by GPS. Radar interferometry succeeded in measuring the pattern of coseismic displacements of the magnitude 7.3 Landers earthquake of 1992 [*Massonnet et al.*, 1993]. The current resolution of vertical displacement of 30 mm, however, is much less than that of conventional techniques. The promise of this technique is that measurements conducted by a dedicated satellite could bring under surveillance hundreds of volcanoes every few weeks.

Almost all efforts in volcano geodesy have been directed at subaerial volcanoes, even though most volcanic activity occurs along the mid-ocean ridge system. Proposed techniques for working under the sea are natural extensions of successful techniques used on land, such as development of an ocean bottom tiltmeter [*Staudigel et al.*, 1991] and measurement of horizontal distances across the ridge [*Chadwick et al.*, 1990; *Spiess et al.*, 1994]. Because the tops of magma reservoirs are within 1–2 km of the seafloor and less than 3 km wide [*Detrick et al.*, 1987; *Collier and Sinha*, 1992], instruments will have to be placed within a few kilometers of the axis of the ridge system. If the 1978 event at Asal-Goubbet is typical of events along the undersea portion of the mid-ocean ridge system, the instruments must be spaced about every 5 km along a 500-km length of the ridge system in order to have a reasonable chance of detecting an event in 5 years [*Stein et al.*, 1991]. The probability of detecting an event, however, rises dramatically with spreading rate. Also, a higher chance of detection is possible if initial efforts are concentrated at seamounts along the mid-ocean ridge system, which may be loci of activity. For example, an ocean bottom pressure sensor that was operated for 15 months on Axial seamount, Juan de Fuca Ridge, recorded a 150-mm subsidence coincident with a sudden change in water temperature and an increase in speed of ocean bottom currents, activity consistent with a sudden outburst of confined hot seawater within the oceanic crust [*Fox*, 1990].

The purpose of volcano geodesy has advanced far beyond the initial objective of recording the rise and fall of the surface as magma rose beneath, then erupted from, a volcano. The results are beginning to yield estimates of magma supply rate to volcanoes and to record the size and growth rate of intrusions. In some cases the rate of displacement has been used to predict the start of an eruption, though this goal has not been achieved in every case. In the future, such measurements may achieve the equally important goal of determining when an eruption is about to end.

ACKNOWLEDGMENTS. Hundreds of people have contributed to the growing science of volcano geodesy. Unfortunately, only a few can be acknowledged here. Robert Decker serves as a mentor who continues to pose questions about the nature of ground displacement around active volcanoes. We have tried to answer many of his questions in this paper. The professional acumen of Don Swanson and Richard Fiske impressed us to delve farther into the field. In part, this paper was inspired by their accomplishments and skepticism. The capable staffs of the Hawaiian and Casacades Volcano Observatories taught both of us how to conduct field measurements and set us to wondering what was the cause of the displacement patterns that were revealed. This paper has benefited greatly from the thorough comments and additions suggested by John B. Murray and Paul Davis.

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